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PALEOECOLOGY OF THE MIDDLE PLIOCENE DEER LODGE LOCAL FAUNA, WESTERN MONTANA

BY R. L. KONIZESKI

ABSTRACT

Deer Lodge Valley lies to the northwest of Butte, Montana. It is an early Tertiary structural trough, partially filled with late Tertiary alluvium. A mammalian vertebrate fauna collected from middle Pliocene sediments contains ecological analogues of the modern Deer Lodge ecosystem.[1] Sediment distribution suggests similarity of the middle Pliocene pattern of precipitation to the modern pattern. Middle Pliocene habitat zones, species distribution, and population density are considered analogous to those of the modern ecosystem. Recent approximation of the pre-Pleistocene ecological situation testifies to the presence and effectiveness of regional controls that maintained, or preserved the potential to maintain, strong selection pressures during the described geological interval. The replacement of numerous middle Pliocene species by immigrant forms indicates that environmental relationships have played a major role in the reorganization of the valley's mammalian fauna.

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[1] The stratigraphic nomenclature used in this paper follows that of Wood *et al.* (1941).

Introduction

General Statement

This paper attempts to supplement a geological investigation with biological interpretations. The ecosystem has been defined (Allee *et al.*, 1949, p. 695) as the interacting environmental and biotic systems. The present investigation is restricted to an analysis of the physical environments and mammalian components of a paleoecosystem. It is developed through an integrated study of the modern Deer Lodge Valley ecosystem and regional geology, especially the middle Pliocene sediments and fossils.

Some of the more important considerations requisite for the successful investigation of the paleoecosystem include:

(1) Availability of fossil materials—the fossil assemblage must be adequate for accurate interpretation of the original species representation.
(2) Availability of modern samples—a knowledge of the modern ecosystem existing within the same area and in the natural state is prerequisite for comparative purposes.
(3) Interbiotic relationships—morphological similarities of the fossil to modern species must allow accurate description of the ecological relationships of the former by use of analogies drawn from the latter.
(4) Geology—the geological setting must be such that significant ecological interpretations can be drawn from it.
(5) Definition—the paleoecosystem must have precise temporal and areal boundaries.

Choice of the area of investigation was dictated by these considerations.

Location

Deer Lodge Valley heads on the western slope of the Continental Divide near Butte, Lat. 46° N., Long. 112° 30′ W., in Silverbow County, Montana. It trends north and west to Drummond, Lat. 46° 40′ N., Long. 113° W., in Powell County, Montana (Fig. 1). The town of Opportunity lies approximately 5000 feet above sea level, Deer Lodge 4500 feet, Garrison 4300 feet, and Drummond 3800 feet.

The rugged Flint Creek Range bounds the valley on the west. From east to west it is composed of Cretaceous andesitic clastic rocks, Quadrant quartzite, Madison limestone, early Tertiary granitic rocks, and Algonquian rocks. The Garnet Range and Deer Lodge Mountains bound the valley on the east. Between Garrison and Silverbow the northeast valley walls are dominated by andesitic flows and the southeast by the Warm Springs quartz monzonite batholith (Robertson, 1953, p. 5).

High terraces slope from the mountains toward the valley center where they rise above low fringing terraces that bound the inner Deer Lodge Valley (Fig. 2). The paleoecosystem under consideration is limited to the area between Opportunity and Garrison as bounded by the lateral extremities of the high terraces. It is restricted to the time involved in the deposition of the visible middle Pliocene sediments of the terraces.

Previous Work

In the late nineteenth and early twentieth centuries, E. Douglass of the Carnegie Museum, Pittsburgh, collected vertebrate fossils throughout the area. His publications, especially those of 1902, 1903, and 1907, have been an invaluable aid to this work. Since 1941 Dr. J. Le Roy Kay of the Carnegie Museum has devoted considerable time to the collection of Deer Lodge Valley vertebrate fossils. Many of his specimens are unprepared, but the rest have been studied by the author and agree with the conclusions reached herein. Weed (1912) published on the geology of the eastern portions of the area, especially on the Butte-Silverbow district. Billingsley (1915) presents an excellent discussion of the regional geology. Pardee (1950; 1951) and Alden (1953) discuss the physiography and history of western Montana.

Acknowledgments

This paper is based on two summers of field work, 1953–1954, financed by the Department of Geology, University of Chicago.

The writer wishes to express his gratitude to Drs. E. C. Olson, R. L. Miller, the late L. Horberg, J. C. Jamieson, and J. M. Weller, of the University of Chicago, to Dr. J. R. Beer-

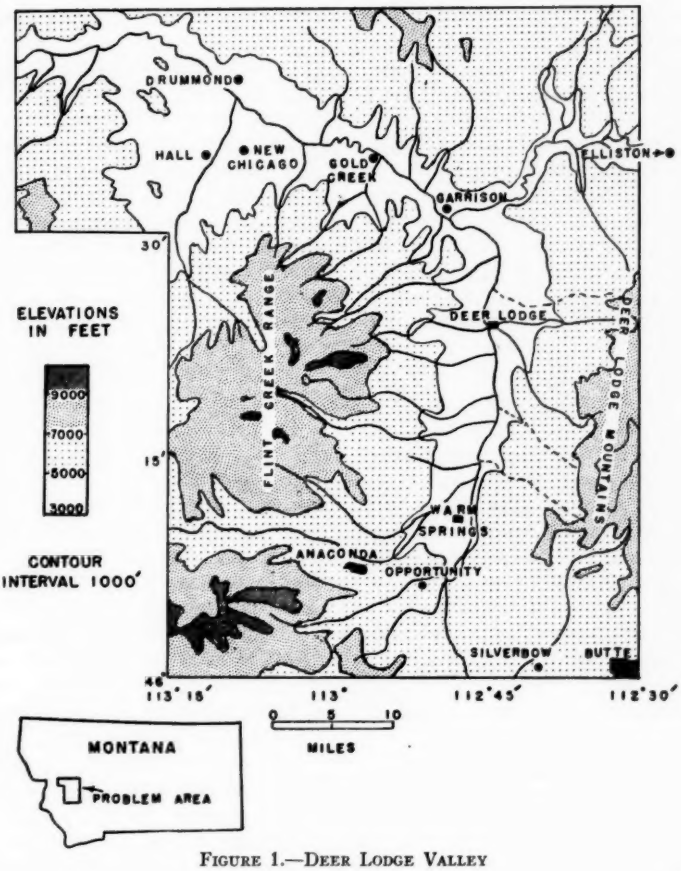


FIGURE 1.—DEER LODGE VALLEY

bower of Lafayette College, and Dr. J. Clark of Princeton University for their helpful suggestions and criticisms throughout the preparation of the paper. Drs. K. P. McLaughlin and F. Honkola of Montana State University initiated the author's interest in the Deer Lodge Pliocene vertebrates; Dr. P. Wright aided in the review of the modern Deer Lodge vertebrate fauna.

Dr. J. Le Roy Kay of the Carnegie Museum gave the writer permission to study his Deer Lodge collections and aided him in the field. Mr. and Mrs. A. Erickson, Mr. and Mrs. K. Rictor, Mr. L. Barton, Mr. D. Young, Mr. and Mrs. W. Tavenner, Mr. F. Ruple, and Mr. W. Konizeski aided in the field work. Mr. J. C. Donohoe and Mr. R. J. Seltin aided in the preparation of fossil specimens and in the review of the final paper.

MODERN ECOSYSTEM[2]

Climate

Western Montana is characterized by rugged north-south trending ranges separated by long narrow valleys. The principal air masses approach the area from the northwest and Pacific Ocean. Most of their moisture falls on the western flanks of the higher ranges. The Flint Creek Range receives up to 30 inches of annual

[2] The modern ecosystem is described only to the extent and in terms applicable to the analysis of the middle Pliocene ecosystem.

precipitation. Its rain shadow extends southeastward over the Deer Lodge Valley.

Mean annual precipitation is approximately 13 inches at Anaconda and 10 inches at Deer Lodge and Drummond. Precipitation in the Deer Lodge Mountains increases northward with elevation because of the lessening influence of the Flint Creek Range rain shadow. Mean annual precipitation at Elliston is approximately 21 inches. Annual snowfall averages 50 inches or more in the higher mountains where it can occur at any time of the year. Snow cover in the valley is seldom continuous through the winter.

Temperature fluctuations are large and rapid because of the alternating influx of warm Chinook winds from the Pacific and cold air masses from east of the Continental Divide. Winters are usually cold; summers are short with warm days and cool nights. From 1911 through 1940 the average January temperature at Deer Lodge was 21.6°, the average July temperature 64.6°. Maximum temperature was 108°, minimum −42°. Growing season in the valley is about 100 days (U.S. Dept. of Agriculture year book, 1941, p. 957, 965).

Distribution of Biota and Habitat Zones

Specimens of the middle Pliocene flora are unavailable, and specific analysis of the modern flora is therefore omitted. Several species of pine and fir provide a medium to dense cover on the deeply dissected and relatively well-watered Flint Creek Range slopes. Conversely, the semiarid western slopes of the Deer Lodge Mountains are characterized by open parks with grass and sagebrush cover. The Deer Lodge Mountain ravines and the cross-terrace ravines on both sides of the valley are well to sparsely timbered with conifers, according to the amount and persistence of water supply. A perennial water supply from the semihumid Flint Creek Range permits intensive cultivation of the west-valley terraces. The east-valley terraces are crossed by only two small perennial and several intermittent streams; they support a sparse cover of harsh grasses and are used chiefly for grazing. Several species of broad-leafed trees and water plants grow along the banks of the Clark Fork River. A variety of grasses and woody perennials are found on the open flood plain.

In the middle nineteenth century four major habitat zones presumably existed within the valley in an unbiased natural state,[3] as defined by vegetational and geomorphological controls. They included: a poorly watered terrace zone; a flood-plain zone including well-watered grazing and, to a limited extent, browsing areas; a riparian zone including the stream banks and channels; and a coniferous cross-terrace ravine zone (Fig. 7). Other zones are believed to have been present but are not described, since their Pliocene analogues cannot be analyzed.

Tables 1 and 2 list the inferred middle nineteenth-century Deer Lodge Valley mammalian fauna.[4] They include generalizations of relative body size, population densities, and food and zonal preferences. Only gross and perhaps misleading values are applied because of limited application to the analysis of the Pliocene ecosystem and because of errors inherent in the subjective evaluation of a fauna of a century ago.

Approximately 49 species of mammals are believed to have lived within the valley 100 years ago. The herbivore-carnivore species ratio was probably about 2 to 1. Population density of the latter was necessarily small compared to that of their prey. Many individuals doubtless ranged outside the valley; many were occasional or annual visitors.

Thirty-five species are listed as unlikely constituents of a conjectural future fossil assemblage; positive criteria are: very small size, low population density, and nonfossorial or arboreal habit. Fourteen species are listed as probable constituents of a conjectural future fossil assemblage; positive criteria are: large body size, high population density, and fossorial habit. Most of the species listed as probable constituents of a future fossil assemblage were multizonal but exhibited a preference for one. The number of species was approximately equal in the terrace, flood-plain, and coniferous

[3] The Indians introduced some bias but it is negligible to the purpose of this paper.

[4] Middle nineteenth-century distribution is based on conversations with Dr. P. Wright of Montana State University and with early settlers, on inferences drawn from the present distribution (Wright, 1951; Burt, 1952; Anthony, 1928), and on personal observation.

zones and lower in the riparian zone. Population density was relatively high in the terrace and flood-plain zones and low in the coniferous and riparian zones.

maximum aeration and ground-water action. Solutional and bacteriological activities apparently destroy most of the remains of the very small animals and plants. Although several

TABLE 1.—EARLY NINETEENTH-CENTURY DEER LODGE VALLEY MAMMALIAN FAUNA

Size, food, and habitat-zone preferences, and relative population densities of animals that because of size, numbers or ecological relationships might be expected to be well or moderately well represented in a conjectural future fossil assemblage.

Animal	Food habits	Population density (d)	Habitat zones							
			Terrace		Flood plain		Riparian		Coniferous	
			p	dp	p	dp	p	dp	p	dp
Pronghorn	large grazer	2	9	18	7	14	..	..	..	..
Bison	large grazer	2	6	12	8	16	..	..	2	4
Rocky Mountain sheep	large grazer	1	8	8	8	8	..	..	..	..
Whitetail jack rabbit	small grazer	3	8	24	6	18	..	..	2	6
Cottontail rabbit	small grazer	3	5	15	6	18	..	..	5	15
Mule deer	large browser	2	3	6	5	10	..	..	8	16
Whitetail deer	large browser	1	3	3	8	8	..	..	5	5
Elk	large browser	1	6	6	7	7	..	..	3	3
Moose	large aquatic plant eater	1	..	..	7	7	6	6	3	3
Muskrat	small aquatic omnivore	2	..	..	..	..	16	32	..	..
Beaver	small aquatic bark eater	1	..	..	2	2	12	12	2	2
Pocket gopher	very small fossorial granivore	3	5	15	8	24	..	..	3	9
Ground squirrel	very small semifossorial granivore	3	6	18	6	18	..	..	4	12
Ground squirrel	very small semifossorial granivore	3	6	18	6	18	..	..	4	12
Total = 14 species		s	11		13		3		11	
		t (dp)	133		174		50		91	
		% t (dp)	29%		39%		11%		20%	

d—total species population density weighted from 1 (low) to 3 (high).
p—species habitat-zone preference weighted from 0 (none) to 16 (total).
dp—species population density within a particular habitat zone.
s—species representation, *ie.*, number of species occurring in a particular habitat zone.
t—habitat-zone population density relative to total population density.
% t (dp)—percentage of habitat-zone population density relative to total population density.

Aspects of Preservation

Partially buried specimens of the present fauna include numerous skeletal elements of the large grazers (cow, sheep, horse, pronghorn, and deer), multitudinous fragments of rabbit skeletons, and a few fragments of bear and pocket-gopher skeletons. Most of the matrix materials are derived from the high terraces. Lack of consolidation of the sediments allows unsuccessful attempts were made, extensive screening in favored areas should result in the recovery of at least a few specimens (Hibbard, 1941, p. 80).

In the east-valley areas maceration of the carcasses and scattering of the skeletal elements generally precede burial in cross-terrace, slope-wash, and intermittent-stream deposits. In a few places local concentrations occur in eddy

TABLE 2.—EARLY NINETEENTH-CENTURY DEER LODGE VALLEY MAMMALIAN FAUNA

Size and food habits of animals that because of size and/or ecological relationships might be expected to be absent or poorly represented in a conjectural future fossil assemblage.

Animal	Food habits	
Flying squirrel	very small arboreal herbivore	
Porcupine	small semiarboreal herbivore	
Red squirrel Meadow mouse (4 sp.) Red backed mouse Jumping mouse White-footed mouse Packrat Pika Golden-mantled marmot Chipmunk (2 sp.) Lemming mouse	very small nonfossorial herbivore	
Shrew (3 sp.) Weasel (2 sp.)	very small nonfossorial carnivore	
Black bear Grizzly bear	large wide-ranging omnivore	population density relatively low
Cougar Wolf	large wide-ranging carnivore	
Coyote Lynx (2 sp.) Fox Otter Wolverine	small wide-ranging carnivore	
Mink Marten Badger	small carnivore	
Skunk	small omnivore	

Total = 35 species
Sum total = 49 species
40 genera
4 omnivores
16 carnivores
29 herbivores

deposits. A few weathered specimens occur on both east- and west-valley terrace tops. Burial has resulted from settling of the heavy bone into the relatively light sands and soils. The west-valley cross-terrace ravines are areas of active erosion and consequently of minimum preservation of faunal elements. Two partially buried skeletons lie in a west-valley ephemeral pond south of Sage Creek. Most flood-plain areas are under extensive cultivation. The modern stream deposits do not lend themselves to prospecting for vertebrate remains.

PHYSICAL GEOLOGY

General Statement

The Deer Lodge Valley Tertiary sediments are as yet undescribed, and the available data are insufficient for their correlation with the known Rocky Mountain section. Geological relationships are considered here only to the extent to which they can provide a temporal framework of physical processes and events, within and to which the paleoecosystem can be oriented.

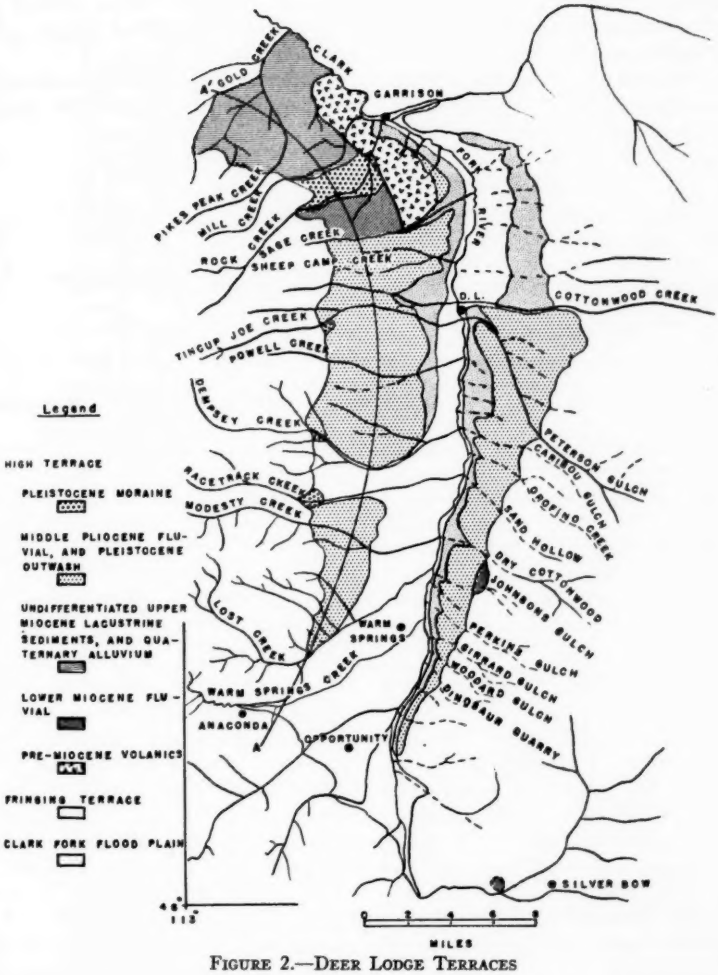


FIGURE 2.—DEER LODGE TERRACES

East-Valley High Terrace

The east-valley high terrace is approximately 3 miles wide at its northern terminus, Cottonwood Creek. It narrows southward and ends opposite Opportunity (Fig. 2). Height of the flood-plain high-terrace scarp remains approximately constant throughout its north-south extent. Side gulches and frontal truncations afford numerous outcrops.

The high-terrace surface dips from 3° to 6° valleyward and northward. Erosional subterraces dip toward controlling tributaries and the Clark Fork River. Cottonwood and Peterson creeks are the only perennial cross-terrace streams. Moderate southeast-valley uplift is reflected by northward lateral migration of the southern cross-terrace tributary ravines (Fig. 3).

With three exceptions, most east-valley high-terrace sediments: are of middle Pliocene Hemphillian age as indicated by the included fossils (table 4); are unconsolidated and include varying amounts of fine volcanic ash and dust;

and excepting the volcanics, have an upslope valley-wall source area. Regional bedding approaches the strike and exceeds by a few degrees the tip of the high-terrace surface. Andesitic rocks parallel the high terrace along the northern one-third, the Warm Springs batholith along the southern two-thirds, of its length. Throughout the batholith-terrace transition area, quartz monzonite is disintegrating and moving via creep, slope, and intermittent-stream wash, to the high-terrace surface. In vertical section the middle and southern terraces consist of approximately 350 feet of alternating loose to semi-resistant beds of sands and grits of similar derivation. To the north along the Cottonwood Creek scarp 100 feet of loose poorly washed stream sands is exposed, and 75 feet is exposed at the Orofino Creek outcrops. Included fragments of andesite were traced to an east-valley-wall source.

In those areas most distant from the mountain front—in the west Peterson Creek to Orofino Creek areas—the sands and grits give way to finely bedded brown unconsolidated silts with dip and strike consistently approximating that of the modern Clark Fork flood plain. Their composition averages 90–95 per cent fragments of volcanic glass, and their coloration is a result of decomposition of incorporated rock and mineral fragments. Five sieve analyses by weight averaged 100 per cent less than ½ mm, 95 per cent less than ½ mm, and 75 per cent less than ⅛ mm.

Discontinuous deposits of horizontally bedded gray tuffs crop out along the high-terrace scarp south of Dry Cottonwood Creek. Five sieve analyses by weight averaged 100 per cent less than ¼ mm, 75 per cent less than ⅛ mm, and 40 per cent less than 1/16 mm. These tuffs are composed of unweathered volcanic glass shards. Spectrographic analyses of single specimens of tuffs and brown silts indicate similar chemical composition (Table 3).

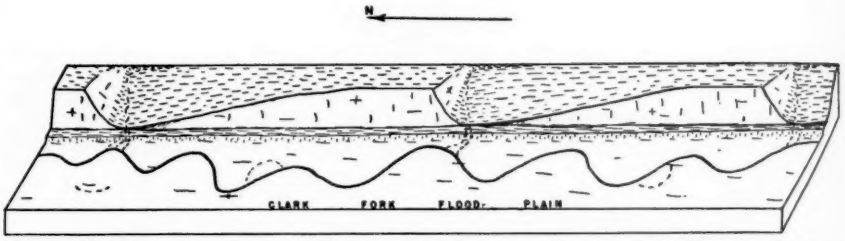


FIGURE 3.—NORTHWARD LATERAL MIGRATION OF SOUTHEAST-VALLEY CROSS-TERRACE RAVINES

TABLE 3.—CHEMICAL COMPOSITION OF GRAY TUFFS AND BROWN SILTS
(In weight per cent)

	SiO_2	Fe	MnO	CaO	MgO	TiO_2	Al_2O_3	Na	K
Gray tuff	72	2.5	0.085	2.60	~1	0.60	14.0	1.65	3.65
Brown silt	80	1.4	0.055	1.20	<1	0.45	12.0	2.70	6.10
Error (±)	3	0.1	0.005	0.15		0.04	0.5	0.05	0.15

In most instances 1–3 inches of fine-bedded white clays underlie the tuffs and overlie several feet of rounded channel conglomerates. These relationships indicate a transition from a stream channel to a lacustrine (probably isolated oxbow) environment of deposition.

Bedded remnants of pre-middle Pliocene breccias crop out on the side of the Warm Spring batholith at the head of Johnsons Gulch. They are gray, well indurated, and approximate the dip and strike of the Pliocene sediments which in some areas directly overlie them. No identifiable fossils were found.

West-Valley High Terrace

The west-valley high terrace averages 3-4 miles wide and extends, with interruptions,

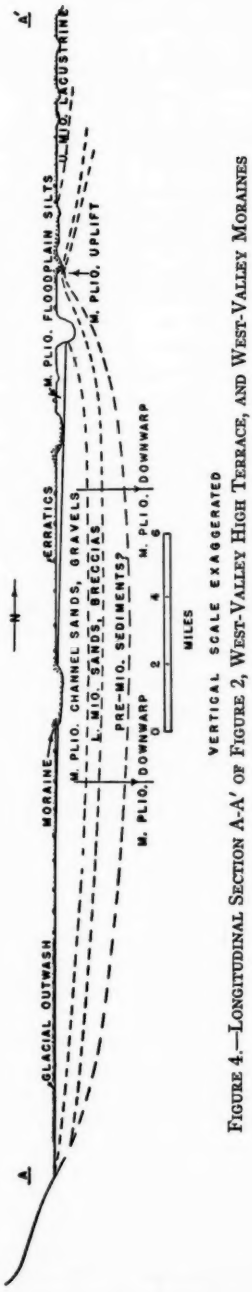


FIGURE 4.—LONGITUDINAL SECTION A-A' OF FIGURE 2, WEST-VALLEY HIGH TERRACE, AND WEST-VALLEY MORAINES

from Lost Creek to Gold Creek (Fig. 2). Its surface merges with the Clark Fork flood plain to the south and rises gradually above it to the north (Fig. 4). Two hundred and fifty to 300 feet of sediments are exposed near the mouth of Sage Creek. Average valleyward dip is 1°–3°, and the longitudinal dip approaches 0° at the terrace scarp.

Glacial outwash caps the high terrace south of the Racetrack Creek scarp. Large deeply weathered glacial erratics are scattered over much of its surface. Flint Creek Range valley moraines spill out along the mountain—high-terrace line of juncture (Fig. 2). Eleven perennial streams head in the Flint Creek Range and cross the high terrace to join the Clark Fork River. Numerous outcrops in their valleys reveal the structure and lithology of the unconsolidated terrace sediments.

The high-terrace surface bevels lower Miocene, upper Miocene, and middle Pliocene sediments (Fig. 4). The visible section reaches a maximum thickness of 300 feet of unconsolidated middle Pliocene sediments at the Sage Creek scarp and wedges out to the south near Lost Creek. The lower 150 feet of section consists of coarse well-washed channel sands and subangular gravels up to 6 inches in diameter. Several anastomosing channels are traceable by their resistance to weathering because of the cementing properties of incorporated manganese precipitates. Regional bedding approximates the dip and strike of the neighboring Clark Fork flood plain.

One hundred and fifty to 200 feet of brown unconsolidated silts overlie the channel deposits at Sage Creek. They thicken westward to a maximum of 300 feet and wedge out to the south near Modesty Creek. Outliers occur along the western terminus of the terrace as far north as Mill Creek. These silts are petrographically indistinguishable from the east-valley (Peterson Creek to Orofino Creek) brown silts with which they were apparently originally continuous. Their lithology, geographical distribution, and fossil content suggest a flood-plain environment of deposition. Ten feet of gray volcanic dust crops out immediately south of Tincup Joe Creek. It is petrographically indistinguishable from that previously described from the southeast-valley areas.

An early Miocene fossil assemblage was collected from sediments in the area bounded by the West Garrison volcanics and the Flint Creek Range, and by Sage and Rock creeks (Fig. 3). The sediments dip up to 30° to the southeast and are overlain by undisturbed middle Pliocene sediments along the east-west Sage Creek scarp. The section includes 100 feet of well-indurated breccias similar to those in the east-valley Johnsons Gulch area. Correlation on the basis of lithology would probably fail as the source rocks differ widely in character.

The area bounded by Rock and Mill creeks, the mountain front, and West Garrison volcanics is covered by glacial debris. Remnants of upper Miocene sediments extend from Mill Creek to Flint Creek. The upper section includes fine-bedded lacustrine clays, underlain in the lower Flint Creek section by volcanic ash, fresh-water limestones, clays, and some intercalated channel sands. Bedding is approximately horizontal from Flint Creek to Pikes Peak Creek. Dip and strike are variable in the area south of Pikes Peak Creek, averaging 1°–5° to the northwest. The Flint Creek faunal assemblage collected by Douglass (1902, p. 273; 1903, p. 153) includes numerous late Miocene mammals as well as fish and aquatic gastropods.

Clark Fork Flood Plain

The Clark Fork flood plain averages 3 miles in width between fringing terraces (Fig. 2) and dips north at 23 feet to the mile. The towns of Garrison and Opportunity mark its north and south ends. Meanders of the Clark Fork River continue through Garrison Cut where they are superimposed on Tertiary volcanic rocks (Pardee, 1951, p. 80). One to 10 feet of soil covers most of the plain. Several mid-valley gravel pits expose up to 30 feet of well-washed Pleistocene to Recent sands and gravels. They extend downward to an unknown depth and presumably overlie Pliocene and or older Tertiary sediments. Billingsley (1915, p. 37) shows a drill hole near Opportunity as terminating in the "Boulder batholith" complex. The writer has been unable to locate the records of this operation but has estimated a 1500–2000 foot depth of sediments by extending the known valley relief.

Tertiary Sediments from Gold Creek through Drummond

Dissection of the Gold Creek to Drummond terrace surfaces and erosion of the unconsolidated Tertiary sediments have proceeded much further than in the south Garrison areas. Older (upper Oligocene?) sediments are exposed, while lower Miocene and Pliocene sediments appear to be absent. Upper Miocene sediments lie unconformably on (upper Oligocene?) sediments at the Flint Creek New Chicago cemetery, 3 miles east of Hall (Fig. 1). Three miles southeast of Drummond (upper Oligocene?) sediments crop out at equal elevation and 2 miles across valley from the cemetery outcrops (Douglass, 1902, p. 238). All are valley-wall features as opposed to numerous low, mid-alluvial terraces of Pleistocene age. The Oligocene? sediments are composed of brown to gray clays with intercalated beds of volcanic dust and a few large lime concretions.

The Clark Fork River flows northwest from Drummond to cut through mountains of pre-Tertiary rocks with valley-wall relief up to 2000 feet and elevations up to 6000 feet.

Middle Pliocene Deer Lodge Local Fauna

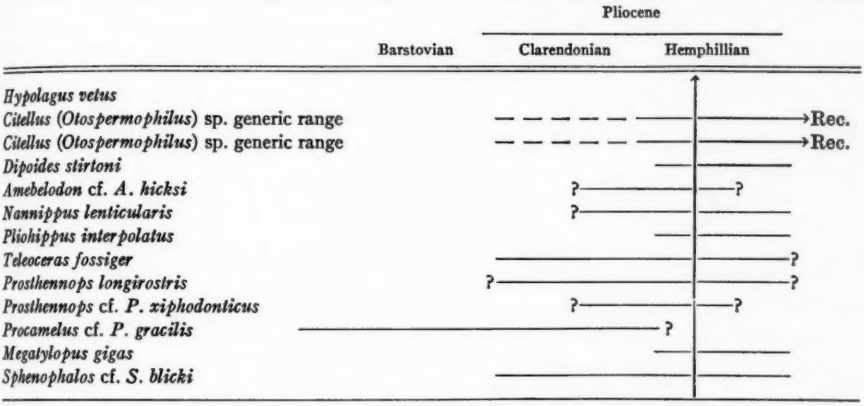
The middle Pliocene fossil assemblage includes 111 specimens ranging from three partial skeletons to single isolated teeth. It represents 11 genera and 13 species (Table 4). Species and specimen distribution, matrix relationships, specimen orientation, and state of preservation have been carefully scrutinized, inasmuch as they provide a basis for analysis of the paleoecosystem.

Most of the specimens were recovered from the east-valley unconsolidated grits. These include representatives of 12 of the 13 species, excluding the beaver, *Dipoides stirtoni*. The fossils usually occurred as isolated fragments; a few local concentrations have been interpreted as eddy deposits. Contributing evidence was the incorporation of fist-sized cobbles and the separation of the deposits by pebble lenses from the underlying sands.

Similarity of the east-valley middle Pliocene

specimens to the products of modern burial suggests analogous processes of preservation—*ie.*, the animals died on the terrace slopes and were buried in slope wash or tributary fill. are among the commonest of the modern specimens undergoing burial. The negative ratio implies that they were in a minority during middle Pliocene time.

TABLE 4.—MIDDLE PLIOCENE DEER LODGE LOCAL FAUNA SPECIES LIST AND STRATIGRAPHIC RANGE



Scattering and fragmentation of the preserved specimens indicate a considerable time lapse between death and burial. A consistent lack of specimen halos[5] is in accordance with these conditions.

Despite extensive screening, the only very small forms found are single specimens of two species of *Citellus* (*Otospermophilus*) sp. and of *Hypolagus annectans*. This discrepancy, in contrast to the relatively abundant representation of the larger species, is analogous to the modern situation. It probably results in part from destruction of the very small animals by the mechanical and chemical processes inherent in the described processes of preservation. *Citellus*, a ground squirrel, is a fossorial type. Both specimens occurred as partially crushed skeletons, and the implication is that these individuals died in their burrows and their skeletons were preserved there. The rabbit, *Hypolagus*, is represented by a single fragment of lower jaw found in a local concentration of coarse aggregates including fossil fragments of larger animals. Elements of rabbit skeletons

[5] Halos are the hardened and discolored areas of matrix directly in contact with, and surrounding, some fossils. This phenomenon is believed to be a result of chemical activities stemming from the decay and maceration of the carcass during burial.

Outcrops are fewer and less extensive in the west-valley channel sands than in the east-valley sediments. Fossils are more numerous per given area but are restricted to six of the larger species and a beaver, *Dipoides stirtoni*. All the specimens show the effects of extreme abrasion. Two species of horse (*Nannippus lenticularis* and *Pliohippus interpolatus*), an ancestral pronghorn (*Sphenophalos* cf. *S. blicki*), and a peccary (*Prosthennops longirostris*) are represented by half a dozen isolated teeth. Numerous large fragments were found of the hippopotamus-like rhinoceros (*Teleoceras fossiger*) and a shovel-tusked mastodon (*Amebelodon* cf. *A. hicksi*). A complete skull of the former and a partial articulated skeleton of the latter were recovered from the Dempsey Creek gravel pit. In view of the robust nature of the large camel (*Megatylopus gigas*) its omission from the west-valley channel sands is probably due to habitat-zone preference rather than to preservational bias.

Exposures of the flood-plain silts are numerous, but fossils are few. Five teeth of *Sphenophalos* cf. *S. blicki* and a skull, an upper molar, and a metapodial of *Nannippus lenticularis* were collected as isolated specimens. Several proboscidean phalangeal elements were

left in the field. The relatively unworn and unfragmented condition of these specimens agrees with the hypothesis of a flood plain environment of deposition.

Species stratigraphic range and specimen distribution of the Deer Lodge local fauna indicate stratigraphic equivalence of the east-to-west-valley sediments (Table 4). Differences in matrix and specimen distribution may be related to differences in the environments of deposition, species distribution, and inadequate samples.

A few concentrations of silica have been interpreted as precipitate fillings of root cavities. They are apparently the sole representatives of the middle Pliocene Deer Lodge Valley flora. Lack of plant preservation is in harmony with the modern situation and suggests a similarity of environment of deposition.

Class MAMMALIA

Order LAGAMORPHA

Family LEPORIDAE

Hypolagus vetus (Kellogg)

C.N.H.M. UM. 556, partial left mandible with P_3-M_3. Collected from middle Pliocene (Hemphillian) grits, Johnsons Gulch quarry, south side of Johnsons Gulch, 2 miles above mouth, Deer Lodge County, Montana.

Order RODENTIA

Family SCIURIDAE

Citellus (*Otospermophilus*) sp.

C.N.H.M. UM. 557, badly broken skull with skeletal fragments. Dentition complete but badly worn. Collected from middle Pliocene (Hemphillian) sands at high-terrace scarp midway between Sand Hollow Gulch and Dry Cottonwood Creek, Deer Lodge County, Montana.

C.N.H.M. UM. 558, complete skull and jaws with skeletal fragments. Collected from middle Pliocene (Hemphillian) sands at high-terrace scarp midway between Girrard and Woodard gulches, Deer Lodge County, Montana. Size differences on the order of 1 to 2 and differences in skeletal and dental characters show that this specimen is specifically different from C.N.H.M. UM. 557.

Family CASTORIDAE

Dipoides stirtoni (Wilson)

M.S.U. 2, partial right ramus with complete cheek series. Collected by Dr. K. De Greene from Dempsey Creek gravel pit, Powell County, Montana.

Order PROBOSCIDES

Family GOMPHOTHERIIDAE

Amebelodon cf. *A. hicksi* (Cook)

M.S.U. 1, partial skeleton with complete lower jaws and skull fragments. Dempsey Creek gravel pit, Powell County, Montana.

Order PERISSODACTYLA

Family EQUIDAE

Nannippus lenticularis (Cope)

C.N.H.M. UM. 560, 561, 562, 3 partial palates and Maxillaries. Middle Pliocene (Hemphillian), Johnsons Gulch quarry, south side of Johnsons Gulch, 2 miles above mouth, Deer Lodge County, Montana.

C.N.H.M. UM. 563, partial palate with right maxillary. Horizon and locality as above.

C.N.H.M. UM. 564, 565, 2 pairs of lower jaws. Horizon and locality as above.

C.N.H.M. UM. 566–569, 4 fragments of mandibles. Horizon and locality as above.

C.N.H.M. UM. 570, partial left mandible with 4 incisors and 1 premolar. Horizon and locality as above.

C.N.H.M. UM. 571, left upper and lower tooth series in occlusion. Horizon and locality as above.

C.N.H.M. UM. 572–577, 4, 3, 3, and 2 associated upper molars and 11 isolated teeth, 1 metapodial probably this species. Horizon and locality as above.

C.N.H.M. UM. 578, upper molar. Middle Pliocene (Hemphillian), Dinosaur quarry, Deer Lodge County, Montana.

C.N.H.M. UM. 579, lower molar. Middle Pliocene (Hemphillian) an eighth of a mile west of Dinosaur quarry, Deer Lodge County, Montana.

C.N.H.M. UM. 580, lower molar. Middle Pliocene (Hemphillian), railroad cut east of Opportunity, Deer Lodge County, Montana.

C.N.H.M. UM. 582, upper molar. Middle Pliocene (Hemphillian), Deer Lodge County gravel pit, Montana.

C.N.H.M. UM. 583, lower molar. Middle Pliocene (Hemphillian), Dempsey Creek gravel pit, Powell County, Montana.

C.N.H.M. UM. 584, upper molar. Middle Pliocene (Hemphillian), 2 miles west of Clark Fork River midway between Dempsey and Powell creeks Powell County, Montana.

C.N.H.M. UM. 585, upper molar. Middle Pliocene (Hemphillian), north side of Tincup Joe Creek, 1 mile above mouth, Powell County, Montana.

C.N.H.M. UM. 586, a skull; 587, an upper molar; 588, and a metapodial probably this species. Middle Pliocene (Hemphillian), midway between Sage Creek and Tincup Joe Creek, 1 mile west of Clark Fork River, Powell County, Montana.

Pliohippus interpolatus (Cope)

C.N.H.M. UM. 589, 590, 2 skulls, crushed along palatal plane and retaining complete dentitions. Middle Pliocene (Hemphillian), Johnsons Gulch quarry, south side of Johnsons Gulch, 2 miles above mouth, Deer Lodge County, Montana.

C.N.H.M. UM. 591–594, 3 lower jaws and 4 isolated lower molars. Horizon and locality as above.

C.N.H.M. UM. 595, 596, an upper molar and a metapodial probably this species. Middle Pliocene (Hemphillian), first side draw, south side of Sand Hollow Gulch, Powell County, Montana.

C.N.H.M. UM. 581, upper molar. Middle Pliocene (Hemphillian), 3 miles above mouth on Modesty Creek, Deer Lodge County, Montana.

Family RHINOCEROTIDAE

Teleoceras fossiger (Cope)

C.N.H.M. UM. 597, left mandible. Middle Pliocene (Hemphillian), south side of Sand Hollow Gulch, half a mile above mouth, Powell County, Montana.

M.S.U. 3, a skull and 3 mandibular fragments. Middle Pliocene (Hemphillian), Dempsey Creek gravel pit, Powell County, Montana.

C.N.H.M. UM. 598, 2 molars. Horizon and locality as above.

C.N.H.M. UM. 599, 1 humerus. Horizon and locality as above.

C.N.H.M. UM. 600, upper molar. Middle Pliocene (Hemphillian), 1 mile up intermediary gulch midway between Johnsons and Perkins gulches, Deer Lodge County, Montana.

The large range of variation in the specimens assigned to this species has made it a catchall for Pliocene Teleoceras specimens. These Deer Lodge specimens are provisionally assigned with the realization that later work may necessitate their reclassification.

Order ARTIODACTYLA

Family TAYASSUIDAE

Prosthennops longirostris (Thorpe)

C.N.H.M. UM. 601, M^3. Middle Pliocene (Hemphillian), Deer Lodge County gravel pit, Deer Lodge County, Montana.

Prosthennops of *P. xiphodonticus* (Barbour)

C.N.H.M. UM. 602, 603, partial skull with left I^3, C^1, P^{2-4}, lower jaws with complete dentition; partial right maxillary with P^{2-4}. Middle Pliocene (Hemphillian), Johnsons Gulch quarry, south side of Johnsons Gulch, 2 miles above mouth, Deer Lodge County, Montana.

Dental formula, 12/3, C1/1, P3/3, M?/3; tooth size and pattern are similar to those given for the type. One exception involves a possible inadvertent reversal of P_2-P_3 of the type (Matthew *in* Barbour 1925, p. 27). Definite identification of the Deer Lodge specimen is reserved until an unworn specimen can be obtained.

Order ARTIODACTYLA

Family CAMELIDAE

Procamelus cf. *P. gracilis* (Leidy)

C.N.H.M. UM. 604, partial palate and maxillaries with left M^{1-3}, right M^{2-3}. Middle Pliocene (Hemphillian), Johnsons Gulch quarry, south side of Johnsons Gulch 2 miles above mouth, Deer Lodge County, Montana.

C.N.H.M. UM. 605, 2 deciduous upper premolars in association and 1 isolated upper premolar. Horizon and locality as above.

C.N.H.M. UM. 606, isolated P_4. Middle Pliocene (Hemphillian), south side of Sand Hollow Gulch, half a mile above mouth, Powell County, Montana.

C.N.H.M. UM. 607, fragment of right mandible with M_2. Horizon and locality as above.

C.N.H.M. UM. 608, P_4-M_1, and M_{2-3}, these specimens possibly associated. Middle Pliocene (Hemphillian), Johnsons Gulch quarry, south side of Johnsons Gulch, 2 miles above mouth, Deer Lodge County, Montana.

Procamelus gracilis (Leidy) is founded on 3 upper premolars. The Deer Lodge specimens do not include upper premolars and are provisionally referred to that species. Tooth dimensions and pattern are similar to the University of California specimen of *Procamelus gracilis*, U.C. 29612. The only observed difference, and one of doubtful significance, is the omission in the Deer Lodge specimen of a prominent accessory lingual cusp on M^1, described by Stirton (1929, p. 297) as occurring in the Fish Lake specimen.

Megatylopus gigas (Matthew and Cook)

C.N.H.M. UM. 609, 610, 627, 628, a pair of lower jaws lacking right incisors, condyles, angles, and parts of the coronoid processes; an isolated P_4; a cervical vertebra; part of a pelvis probably this species. Middle Pliocene (Hemphillian), first tribu-

tary ravine, south side of Sand Hollow Gulch, Powell County, Montana.

Family ANTILOCAPRIDAE

Sphenophalos cf. *S. blicki* (Frick)

C.N.H.M. UM. 611–614, 1 right and 1 left mandible, each with P_4-M_3; 1 right mandible with P_4-M_2, 8 isolated teeth. Middle Pliocene (Hemphillian), Johnsons Gulch Quarry, south side of Johnsons Gulch, 2 miles above mouth, Deer Lodge County, Montana.

C.N.H.M. UM. 615, 616, fragments of right mandible with M_1-M_3; 3 associated deciduous teeth, P_{2-4}. Middle Pliocene (Hemphillian), south side of Sand Hollow Gulch, half a mile above mouth, Powell County, Montana.

C.N.H.M. UM. 617, isolated M_3. Middle Pliocene (Hemphillian), Deer Lodge County gravel pit, Deer Lodge County, Montana.

C.N.H.M. UM. 618, 619, 2 isolated M_3. Middle Pliocene (Hemphillian), midway between Dempsey and Powell creeks, 100 feet west of State prison-farm road, Powell County, Montana.

Despite agreement of tooth pattern and measurements, lack of horn cores prevents definite specific assignment to the San Ildefonso specimen, no. F.A.M. 31689, referred by Frick (1937, p. 495) to this species.

PLEISTOCENE FAUNA

Order RODENTIA

Family CASTORIDAE

Castor californicus (Kellog)

C.N.H.M. UM. 559, right mandible lacking M_3 with badly shattered P_4 and incisor. Collected by Mr. L. Barton from talus at high-terrace scarp midway between Sand Hollow Gulch and Dry Cottonwood Creek, Deer Lodge County, Montana.

Order PERISSODACTYLA

Family EQUIDAE

Equus proversus (Merriam)

C.N.H.M. UM. 625, 626, fragment of maxillary with P^{2-3-4}; fragment of mandible with M_{2-3}. From reworked sands, south side of Sand Hollow Gulch, half a mile above mouth, Powell County, Montana.

EARLY MIOCENE TAVENNER RANCH LOCAL FAUNA

The early Miocene fossil assemblage is strongly biased by the abundance of *Promerycochoerus montanus* in contrast to single specimens each of *Gregormymys douglassi*, *Palaeocastor fossor*, and *Miohippus gemmarosae*. Mineralization has proceeded much further than in the comparatively fragile middle Pliocene specimens.

Order RODENTIA

Family GEOMYDAE

Subfamily ENTOPTYCHINAE

Gregorymys douglassi (Wood)

C.N.H.M. UM. 620, mandibular fragment with M_{1-2}. Lower Miocene (Arikareean) saddle between Sage and Rock creeks, Powell County, Montana. Two other specimens are known. The type was collected by Douglass from the lower Miocene 4 miles north of Divide, Montana; the other was collected by Mook from the lower Miocene of the Silverbow area (Wood, 1936, p. 12)

Family CASTORIDAE

Palaeocaster fossor (Peterson)

C.N.H.M. UM. 621, partial left mandible containing incisor remnants, a deciduous P_4 and well-worn M_{1-2}. Horizon and locality as above.

Order PERISSODACTYLA

Family EQUIDAE

Miohippus cf. *M. gemmarosae* (Osborn)

D.L.H.S. 1, a single unworn M^2. Lower Miocene (Arikareean), saddle between Sage and Rock creeks Powell County, Montana. Collected by Mr. R. M. Austin, it is now in the science collection of the Deer Lodge High School. Association of this specimen with abundant *Promerycochoerus montanus* is interesting, inasmuch as the type occurs in the *Promerycochoerus* zone of the lower Rosebud of South Dakota.

Order ARTIODACTYLA

Family MERYCOIDODONTOIDEA

Promerycochoerus montanus (Cope)

C.N.H.M. UM. 622, 623, an isolated molar; a right maxillary and premaxillary with I^3, C^1-P^4. Lower Miocene (Arikareean) saddle between Sage and Rock creeks, Powell County, Montana.

C.N.H.M. UM. 624, isolated P_4. Lower Miocene (Arikareean), middle Pliocene - lower Miocene unconformity south side of Sage Creek, 2 miles above mouth, Powell County, Montana.

Several skulls and jaws have been collected by local residents from the Tavenner Ranch lower Miocene outcrops. The author left much material

in the field, as it was unnecessary to the present investigation.

dipping lower Miocene sediments at the Sage Creek scarp shows that middle Pliocene dep-

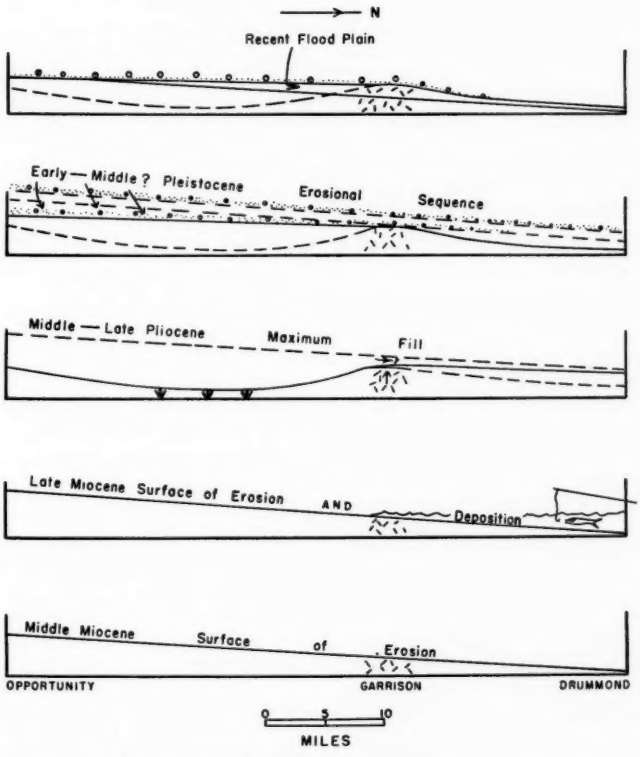


FIGURE 5.—MIDDLE MIOCENE-MIDDLE? PLEISTOCENE EROSIONAL AND DEPOSITIONAL SEQUENCE, DEER LODGE VALLEY, MONTANA

LATE TERTIARY GEOLOGY

A (late Oligocene?) depositional basin existed at Drummond and may have extended into the south Garrison area. Lower Miocene deposits crop out near Silverbow, Garrison, and probably at the head of Johnsons Gulch. Their absence from the Garrison to Drummond area might relate either to their original absence or to their deposition and subsequent removal. Late Miocene lacustrine environments of deposition extended south from Drummond at least to Garrison. Except for local warping, bedding of the lake clays still approaches the horizontal.

The presence of undisturbed middle Pliocene clastics lying unconformably on southeastward-

osition followed and was probably in part coincident with south Garrison downwarping (Fig. 5). Subsidence was confined to the immediate valley areas, as evidenced by the superior elevation of the Johnsons Gulch (lower Miocene?) relative to the middle Pliocene sediments.

Intermittent volcanic dust falls blanketed the valley. Some fell directly into isolated still-water environments to form the gray tuffs. Some fell on the east-valley slopes and was incorporated into the underlying clastics. The dust that fell directly on the flood plain or reached it via lateral transport was carried from the area by the ancient Clark Fork River. The high percentage of dust preserved as

flood-plain deposits shows that an increase in volume partially choked the drainage system during the latter stages of recorded middle Pliocene deposition.

A single specimen each of *Equus proversus* and of *Castor californicus* was collected from the northeastern tributaries were heavily loaded with detritus and subject to much lateral shifting is indicated by the channeling and cross-bedding in their sediments. The resulting east-valley fill built up across valley toward the Flint Creek Range (Fig. 6). The

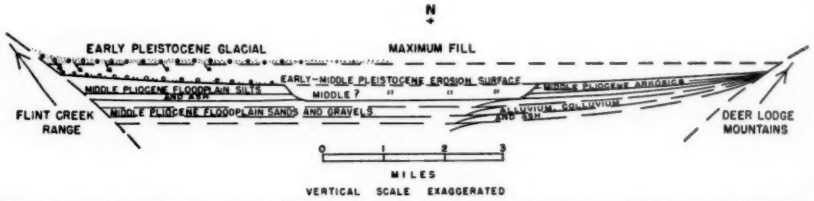


FIGURE 6.—MIDDLE PLIOCENE-MIDDLE? PLEISTOCENE DEPOSITIONAL AND EROSIONAL SEQUENCE THROUGH DEER LODGE VALLEY AT DEER LODGE, MONTANA

east-valley float. Their presence shows that deposition continued into late Pliocene or perhaps Pleistocene time. Lower Pleistocene Flint Creek Range glacial debris was deposited on the west-valley Pliocene sediments (Alden, 1945, p. 80–81) (Fig. 6). Most of the smaller elements and upper? Pliocene sediments were subsequently removed by the Clark Fork River as it cut to the modern high-terrace surface. The larger elements remain stranded on the terrace tops as deeply weathered glacial erratics. Later erosion of the inner Clark Fork valley flood plain was accompanied by the superimposing of stream meanders on resistant Tertiary volcanic rocks through the Garrison Cut. Fringing terraces appear to be late Pleistocene to Recent developments.

MIDDLE PLIOCENE ECOSYSTEM

The estimated total of 350 feet of exposed middle Pliocene sediments contains neither soil zones nor regional unconformities. It probably represents a relatively short period of geologic time. Most of the materials appear to have been derived from semiarid southeast-valley slopes under conditions analogous to the present. Structural relationships suggest southeast-valley, intermittent-stream, and slope wash as the principal agents of erosion. Annual rainfall was greater in the northeastern than southeastern watersheds, as evidenced by the east-valley northward gradation from slope wash to poorly washed stream deposits. That ancient Clark Fork River flowed northward through the middle- and west-valley areas. Its erosional powers were probably augmented by perennial west-valley tributaries inasmuch as it effected the removal of the clay and silt-sized sediments while reworking and depositing the coarse sands and gravels.

Distribution of middle Pliocene sediments suggests a pattern of precipitation similar to the present, with the Flint Creek Range acting as a barrier to moisture-laden northwest winds and the resulting rain shadow developing best to the southeast. The biased, dominantly west-valley distribution of glacial debris indicates that those conditions continued through the Pleistocene to the present.

The hypothesis that middle Pliocene cross-valley relief is comparable to or greater than the present relief is somewhat at variance with the generally held concept of relatively low Rocky Mountain Pliocene relief. It is, however, substantiated by the following evidence: (1) Existence of the middle Pliocene sediments. These must have had highlands for a source and a lowland for a site of deposition. The vertical component is increased in view of the middle Pliocene sediments that were probably deposited before, and Pliocene and Pleistocene sediments that were certainly deposited after the visible middle Pliocene section, plus those sediments carried from the valley during the normal course of deposition. (2) Distribution of the middle Pliocene sediments as described above. (3) Lithology of west valley channel

deposits. The coarse sand and gravel grain size, angularity, and lack of weathering of the individual components are indicative of relatively high relief as well as short distance from source to site of deposition. (4) Regional uniformity and approximation of dip and strike of the middle Pliocene sediments to the modern valley floor. In this light the present vertical discrepancy of several thousands of feet between the undisturbed middle Pliocene sediments and neighboring highlands, if post-depositional, could have been attained only through block faulting. Evidence of post middle Pliocene deformation of such magnitude should however, be clearly visible. The author found none, other than local warping and faulting, and Pardee (1950, p. 402) does not describe any in his comprehensive review of *Late Cenozoic block faulting in western Montana.*

On the basis of late Tertiary floras from California, the Great Basin, and High Plains, Axelrod (1948, p. 127) states, "... those of middle Pliocene age reflect a semi-arid climate, milder and warmer than exists at present in the western United States." The Deer Lodge sediment and faunal distributions are in accordance with these conclusions. In the absence of identifiable structural controls, it appears that the change from valley-wide aggradation to erosion by the Clark Fork River may have resulted from a modern shift to slightly greater annual rainfall. The fauna as a whole is of such composition that its modern analogue would probably be described as south temperate; its collective food habits required mild open winters to insure adequate year-round food supply.

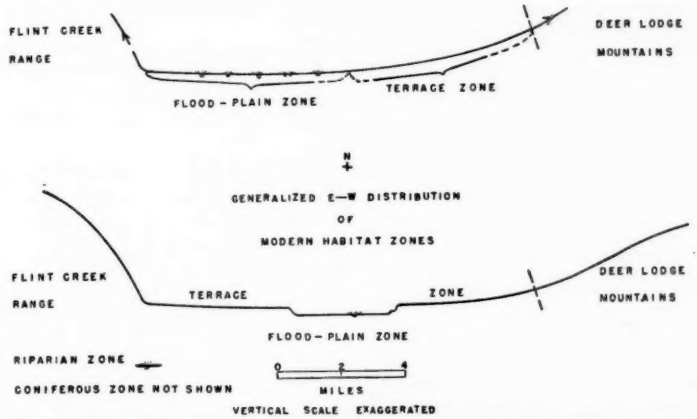


FIGURE 7.—INFERRED EAST-WEST DISTRIBUTION OF MIDDLE PLIOCENE HABITAT ZONES AND GENERALIZED EAST-WEST DISTRIBUTION OF MODERN HABITAT ZONES

The geological and biological evidence indicates the existence of four middle Pliocene Deer Lodge Valley habitat zones. Figure 7 compares their probable cross-valley areal and topographic relationships with those of their modern analogues. The similarity of the two cross-valley profiles is evidenced by the close approach of the dip and strike of the middle Pliocene sediments to that of the modern terrace surfaces.

The poorly watered terrace zone was restricted to the east-valley areas. Comparison of the associated fauna (Table 5) with its modern equivalent (Table 1) indicates a probable ecological similarity of floral elements. Sediment distribution shows that there was continual build-up from the east and shifting of the longitudinal flood-plain axis to the west of its present position. Cross-valley transition from the poorly watered terrace zone to the flood-plain zone appears to have been gradual. The western flood-plain areas probably ended in an abrupt transition to alpine topography. The Deer Lodge Valley valley-wall courses of the perennial tributary streams were

relatively stable in an areal sense, inasmuch as they were cut into bedrock. The immediate areas of emergence onto the main valley floor were therefore also stable and relatively well watered. In this light they were analogous to the modern coniferous zone, although because of the incompleteness of the fossil record it is impossible to know whether conifers were actually present or prominent members in this portion of the middle Pliocene ecosystem. The riparian zone, in an aggrading environment, presumably incorporated a greater area than at present. As the ancient Clark Fork River and its tributaries shifted their flood-plain courses, the sites and attitudes of this habitat zone must have been changing constantly. In the absence of plant specimens it can only be stated that the faunal evidence demands a fairly dense nonconiferous plant cover. Deciduous trees and bushes, grasses, lilies (Merrill, *in* Osborn, 1936, p. 717), and other riparian plants presumably were abundant.

TABLE 5.—MIDDLE PLIOCENE DEER LODGE VALLEY FAUNA

Size and inferred food and habitat-zone preferences and relative population densities of middle Pliocene animals that are represented in the fossil assemblage.

Animal	Food habits	Population density (d)	Habitat zones							
			Terrace		Flood plain		Riparian		Coniferous?	
			p	dp	p	dp	p	dp	p	dp
Pronghorn	large grazer	2	9	18	7	14	..	..	..	..
Horse	large grazer	2	9	18	7	14	..	..	..	..
Horse	large grazer	2	9	18	7	14	..	..	..	..
Camel	large grazer	2	9	18	7	14	..	..	..	..
Camel	large browser-grazer	2	7	14	9	18	..	..	..	..
Rabbit	small grazer	3	8	24	6	24	..	..	2	6
Peccary	large rooter	2	..	..	8	16	4	8	4	8
Peccary	large rooter	2	..	..	8	16	2	4	6	12
Proboscidean	very large aquatic plant eater	1	..	..	6	6	10	10	..	..
Rhinoceros	large aquatic plant eater	1	..	..	6	6	10	10	..	..
Beaver	small bark eater	1	..	..	1	1	14	14	1	1
Ground squirrel	very small semifossorial granivore	3	8	24	2	6	..	..	6	18
Ground squirrel	very small semifossorial granivore	3	8	24	2	6	..	..	6	18
Total = 13 species, 111 specimens		s	8		13		5		6	
		t (dp)	158		156		46		63	
		% (dp)	37%		37%		11%		15%	

d—total species population density weighted from 1 (low) to 3 (high).
p—species habitat zone preference weighted from 0 (none) to 16 (total).
dp—species population density within a particular habitat zone.
s—species representation, *ie.*, number of species occurring in a particular habitat zone.
t (dp)—habitat-zone population density relative to total population density.
% (dp)—percentage of habitat-zone population density relative to total population density.

Table 5 is the analogue of Table 1. The incompleteness of the data tends to invalidate the conclusions. However, the writer believes that Table 5 is justified in that it provides correspondence with a known situation. It lists the middle Pliocene animals represented in the Deer Lodge Valley fossil assemblage, their relative sizes, and inferred food and zonal preferences. The proposed distribution is based on morphology and on the assumption that analogous structures relate to similar ecological situations. It is also based on fossil distribution including aspects of preservation and matrix relationships. It does not rely on the numbers of specimens recovered other than on the presence or absence of a species, nor does

it assume that two related species would necessarily have similar habits.

The Deer Lodge local collection shows a species and specimen paucity in comparison to many recorded middle Pliocene faunas. This is probably because of temporal, areal, environmental, and collecting restrictions. Precise temporal and physical boundaries of many of the larger faunas are difficult to delineate, and the fossil collections are probably long-time accumulations over a large area (Hibbard, 1941, p. 94, 95).

Ten middle Pliocene species of large and small herbivores and 3 species of very small fossorial herbivores are listed, as compared to 11 and 3 at present. Such close numerical agreement is interpreted as evidence of relatively complete species representation of this portion of the middle Pliocene fauna. It suggests the existence of ecological types analogous to those members of the modern assemblage listed as improbable constituents of a conjectural future fossil assemblage. The absence of carnivorous species is probably attendant upon small population, upon a tendency to den outside the immediate valley, and upon inadequate samples. Absence of the very small nonfossorial types is probably due to preservational bias and to biased collecting.

A comparison of the tables indicates analogous species distribution and population density within the two ecosystems—that is, high in the terrace and flood-plain zones compared to the riparian zone. The middle Pliocene coniferous? zone apparently had fewer species and lower population density than that of the present, but the data are too meager to be conclusive.

Discussion

Numerous evidences of multiple Flint Creek Range glaciation exist as tributary-valley moraines and glacial outwash. Although the main Clark Fork Valley was probably never covered with Pleistocene ice, the middle Pliocene habitat zones must have disappeared with the change from a temperate to a subarctic climate. Similar environmental changes probably occurred with each transition from interglacial to glacial conditions. The fauna were forced to retreat to the south, adapt to the new conditions, or die (Matthew, 1915, p. 172).

Of the known middle Pliocene fauna, only the rabbit, beaver, ground squirrel, and antelope are represented in the modern ecosystem. Their Recent recurrence and similarity of the modern to the Pliocene species indicate that these animals retreated to the south and returned with the advent of more equable conditions. Peccaries are now absent from the valley, although they are numerous in the southern United States. If the hypothesis of a middle Pliocene mild warm climate is valid, their absence is probably due to the rigorous Montana winters. Extinction of the camel, rhinoceros, proboscidean, and horse may have resulted from a number of causes including inclement climate. Their places in the ecosystem have been partially filled by immigrant Eurasiatic forms including deer, moose, and elk.

Despite the tentative generalized nature of the analysis of the middle Pliocene Deer Lodge Valley ecosystem, its comparison with the modern ecosystem shows numerous interesting similarities. Habitat zones are essentially analogous both in number and kind, in spite of the strong probability of minor climatic, topographic, and areal differences. Middle Pliocene species have been generally replaced by their Recent ecological equivalents. Zonal representation in terms of total population density and species representation in terms of adaptive types are essentially similar. Such agreement is exceptional in view of the extreme variations known to have occurred during the intervening Pleistocene epoch.

The Recent approximation of middle Pliocene environments testifies to the effectiveness of strong geologically constant controls.[6] These include: (1) the grasses which existed throughout the interim in other localities and probably grew in the valley with the advent of suitable environments during interglacial as well as Recent times; (2) surrounding alpine topography; admittedly undergoing continual change but always present; (3) regional patterns of wind circulation. The replacement of numerous

[6] Includes controls that maintained or preserved the potential to maintain, strong selection pressures during the described geological interval.

middle Pliocene animals by unrelated Eurasiatic ecological analogues indicates that environmental relationships have been more important than genetic relationships in determining the reorganization of the valley's mammalian fauna.

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PALEOZOIC EVOLUTION OF THE GEOSYNCLINAL MARGIN NORTH OF THE SNAKE RIVER PLAIN, IDAHO-MONTANA

By Robert Scholten

ABSTRACT

The hinge belt between the Paleozoic Rocky Mountain geosyncline of central Idaho and the cratonic shelf of southwestern Montana is a critical zone with respect to stratigraphic changes. Structural data and isopach patterns indicate that this zone was recurrently uplifted during the Paleozoic. The Skull Canyon Disturbance during the Cambrian or Early Ordovician gave rise to a sharp angular unconformity truncating Upper Precambrian (Belt) rocks. Additional isolated uplifts were produced in Devonian and Early or Middle Mississippian times.

The hinge is thought to be a zone of weakness in the crust which was especially responsive to early stresses in the geosyncline. This same zone became the site of the greatest deformation during the Laramide orogeny, yielding along a prominent belt of major overthrusts.

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INTRODUCTION

Scope of Problem

Comprehensive reports dealing with the stratigraphy and areal distribution of Paleozoic rocks in central Idaho and adjacent parts of southwestern Montana (Fig. 1) have been published by Umpleby (1913; 1917; Umpleby, Westgate, and Ross, 1930), Ross (1934a; 1934b; 1937; 1947), Sloss and Moritz (1951; Sloss, 1950), Klepper (1951), Gardner *et al.* (1946), and Scholten, Keenmon, and Kupsch (1955). These studies have revealed a paleotectonic framework composed of a geosynclinal basin in central Idaho and a relatively more stable cratonic shelf in southwestern Montana. Ross (1936, p. 372–373) states that the Paleozoic geosyncline was confined to the east of a

positive region which is now the site of the Idaho batholith. The eastern border of the exposed part of the batholith passes northward through the central parts of the Sawtooth and

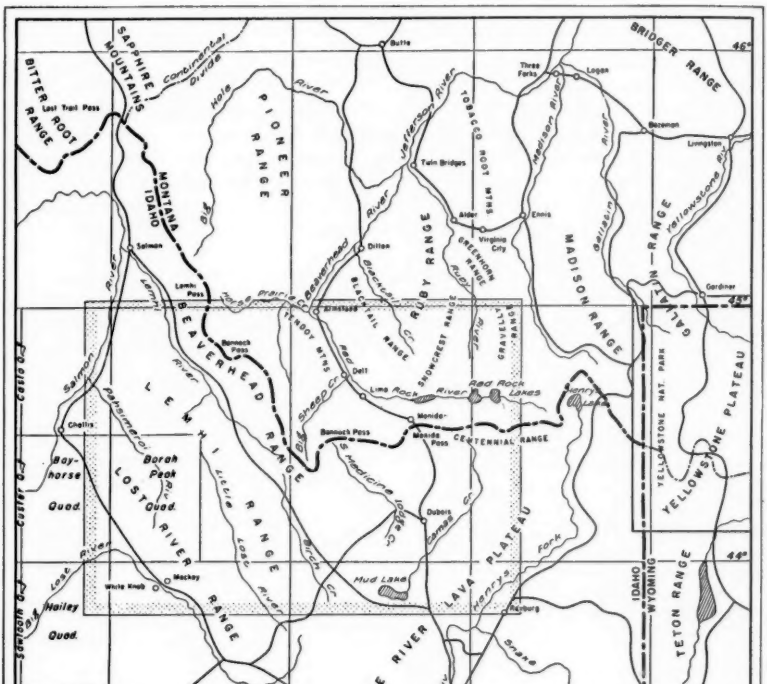

FIGURE 1.—LOCATION MAP
Dotted border outlines area shown on Figures 2–6.

Custer quadrangles and the southwestern part of the Casto quadrangle, just outside the map area covered by Figure 1.

The hinge zone between geosyncline and shelf follows approximately the western border of Montana. This area, which has received comparatively little study, is critical with respect to stratigraphic changes. Field mapping in the southern part of the Beaverhead Range, southeast-central Idaho (Figs. 2–6), during 1954 and 1955 has provided additional information in this critical zone and shed light on several significant departures from regional patterns. The geologic map and a detailed discussion of the geology will be presented in a later paper. The present discussion explores the paleotectonic implications of the areal stratigraphy and its relations to the better-known regions to the east and west. Only the broad outlines of the stratigraphy are presented in an effort to reconstruct the tectonic evolution of this portion of the Paleozoic geosynclinal margin in the northern Rocky Mountains.

Acknowledgments

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The writer is indebted to Dr. C. P. Ross of the United States Geological Survey and Dr. L.

L. Sloss of Northwestern University for making available some unpublished stratigraphic data on central Idaho. They, as well as Mr. T. A. Hendricks of Stanolind Oil and Gas Company, also provided invaluable constructive criticism of the manuscript but should not be considered responsible for the conclusions presented in this paper.

Thanks are also due Dr. Frank M. Swartz of The Pennsylvania State University and Dr. G. A. Cooper of the United States Geological Survey, who identified a number of fossils.

Stratigraphy

Stratigraphic Column

Table 1 shows generalized composite sections of the Paleozoic representative of the central Idaho geosynclinal area and the cratonic shelf of southwestern Montana. All units in both regions are marine, with the probable exception of most of the Permian (?) Casto volcanic rocks.

In the shelf region diastems, major disconformities, and lithologic variety are the rule, and in most places thicknesses do not exceed 1000–1500 feet per stratigraphic system. By comparison, the stratigraphic section in the geosyncline is characterized by much greater thickness and stratigraphic continuity, and vertically the units, on the whole, exhibit greater lithologic uniformity. A general facies change takes place from a western, dominantly clastic and partly volcanic (eugeosynclinal) belt to an eastern, dominantly carbonate-orthoquartzite (miogeosynclinal) belt. In most time-rock units the facies transition occurs roughly across a zone through the Bayhorse quadrangle, slightly west of the Lost River Range (Fig. 1), and in Table 1 this zone is taken as the boundary between the two belts. However, at times miogeosynclinal conditions shifted west (as in the Middle to Late Ordovician), or eugeosynclinal conditions shifted east (as during Early and Late Carboniferous time).

Lateral Variation

General statement.—The detailed lateral variations in the thicknesses of the most important stratigraphic units along the geosynclinal margin and shelf edge are illustrated by the isopach maps (Figs. 2–6). In some parts of the area thickness data are unavailable or known only in a general order of magnitude because of lack of work, absence of outcrops, or questionable correlations, and in such places the positions of the isopach lines have been inferred. However, the regional patterns are sufficiently reliable as a basis for broad paleotectonic interpretation.

Cambrian System.—Middle and Upper Cambrian rocks have been identified in southwestern Montana (Sloss and Moritz, 1951, Fig. 6; Scholten *et al.*, 1955, Pl. 1), where in many places they aggregate more than 1000 feet. Westward and southward the Upper Cambrian disappears, and in the southern Beaverhead Range the entire Cambrian appears to be absent (Fig. 2). In a few exposures quartzites mapped as Belt (Precambrian) may belong to the Middle Cambrian Flathead formation, but in many localities in the southern Beaverhead Range Ordovician or younger rocks directly overlie typical Belt. This area is part of a large region in southeast-central Idaho and southwestern Montana where the Cambrian is absent (Sloss, 1950, Fig. 4; Sloss and Moritz, 1931, Fig. 4a).

On the west slope of the southern Beaverhead Range this hiatus is marked by a sharp angular unconformity between the Ordovician and Precambrian sandstones (Pl. 1), which may be observed in Skull Canyon (Fig. 2) and adjacent canyons to the north and south. The difference in the angle of dip above and below the unconformity is 20°–40°, and the difference in strike is considerable. In most exposures the Belt rocks strike northwest to west and dip north at varying angles, whereas the Ordovician strata depart only slightly from the horizontal. In some places subangular cobbles of Belt sandstone rest on the unconformity. The surface of the unconformity seems to have considerable relief, but this impression may be caused partly by local faulting. Ross (1947, p. 1101; personal communication) has observed a slight angular unconformity at this horizon in the southern Lemhi Range to the west.

Much farther west in central Idaho Ross (1937, p. 12–14 Pl. 1) mapped the Bayhorse dolomite and Garden Creek phyllite, which he provisionally assigned to the Cambrian; he

TABLE 1.—PALEOZOIC STRATIGRAPHIC COLUMN

(Thickness in feet)

Geosyncline of central Idaho SW. of Lemhi Range, Fig. 1 (after Ross, 1934a; 1934b; 1937; 1947, except where credited otherwise)		Cratonic shelf of SW. Montana NE. of Tendoy Range, Fig. 1 (after Sloss and Moritz, 1951; Scholten *et al.*, 1955)
Western belt (mainly eugeosynclinal) through W. Bayhorse quad. and west, Fig. 1	*Eastern belt* (mainly miogeosynclinal) extending slightly W. of Lost River Range, Fig. 1	
PERMIAN		
Casto volcanic rocks 2000–5000 Intermediate to acidic lavas and tuffs; conglomerate at base; minor interbedded limestone (present in and near Casto quadrangle; Permian age not conclusively established)		*Phosphoria formation* 50–650 Limestone, dolomite, black shale, phosphorite, chert, and sandstone
PENNSYLVANIAN		
Wood River formation 8000±	500+	*Quadrant quartzite* 100–1800
Sandstone, fine-grained, impure, quartzitic, locally calcareous; some beds of chert, argillaceous grit, carbonaceous argillite, limestone; basal conglomerate (upper part in Hailey quadrangle Early Permian, according to Bostwick, 1955, p. 944)		Quartzite, light gray to buff, fine- to medium-grained; subordinate cherty dolomite increasing toward top (upper part may be Early Permian)
UPPER MISSISSIPPIAN-LOWER PENNSYLVANIAN		
Milligen formation 3000± Argillite, carbonaceous, black, partly calcareous, thin-bedded, soft, locally hard, siliceous; some impure quartzite, limestone, coal (base may be latest Devonian; top probably Pennsylvanian according to Sloss)	*Brazer limestone* 0–4500? Limestone, medium to dark gray, minor dolomite, abundant chert in nodules and beds; subordinate interbedded units of fine-grained quartzitic sandstone	*Amsden formation* 90–200 Limestone, dolomitic, shaly, with calcareous shale and siltstone, red, buff, gray *Big Snowy group* 0–800 Limestone, shale, dark gray, brown; red shale, siltstone, and gypsum in lower part
LOWER TO MIDDLE MISSISSIPPIAN (Kinderhook-lower Meramec)		
	Milligen formation 1000± Shale, dark gray to black, carbonaceous; some shaly limestone; minor chert (upper beds may be Chester according to Weller *et al.*, 1948, p. 140)	*Madison group* 1000–2000 Limestone, medium to dark gray, with chert layers and nodules; upper part massive (Mission Canyon formation), lower part thin-bedded, shaly (Lodgepole formation); erosional unconformity at top
UPPER DEVONIAN		
	Three Forks limestone 20–350 Limestone, shaly, gray to yellowish brown, thin-bedded; also calcareous shale	*Three Forks formation* 100–200 Shale, siltstone, and limestone, gray to yellowish brown, thin-bedded

TABLE 1.—*Continued*

Geosyncline of central Idaho SW. of Lemhi Range, Fig. 1 (after Ross, 1934a; 1934b; 1937; 1947, except where credited otherwise)		Cratonic shelf of SW. Montana NE. of Tendoy Range, Fig. 1 (after Sloss and Moritz, 1951; Scholten *et al.*, 1955)
Western belt (mainly eugeosynclinal) through W. Bayhorse quad. and west, Fig. 1	*Eastern belt* (mainly miogeosynclinal) extending slightly W. of Lost River Range, Fig. 1	
MIDDLE TO UPPER DEVONIAN		
	Grand View dolomite 1250–2115 Dolomite, well-bedded, fine to medium crystalline, dominantly light to medium gray, partly rusty, sandy *Jefferson dolomite* 700–1150 Dolomite, massive to well-bedded, fine to medium crystalline, mostly dark gray; some buff quartzite	*Jefferson dolomite* 200–500 Dolomite, massive to thin-bedded, fine to medium crystalline, dark gray, buff, partly limy, shaly; some dolomitic shale, mostly in lower part (upper part corresponds to Grand View dolomite of Idaho, according to Sloss, 1954)
MIDDLE SILURIAN		
	Laketown dolomite 325–3000+ Dolomite, finely crystalline, light bluish gray, some pink; interbedded light quartzite and some chert	
Trail Creek formation 500+ Quartzite, light, and argillite, siliceous, calcareous, carbonaceous		
UPPER ORDOVICIAN		
Phi Kappa formation 9400+ Shale and argillite, flinty, dark; sandstone, medium	*Saturday Mountain formation* 3000± Dolomite, shaly, dark; shale, carbonaceous	*Saturday Mountain dolomite* (Fish Haven dolomite of Sloss, 1954) 500–700 Dolomite, poorly bedded, very fine-grained, dark to medium gray; base shaly
MIDDLE TO UPPER ORDOVICIAN		
to fine grained, partly shaly, yellow (present in Hailey quad.)	*Kinnikinic quartzite* 3500± 3000± Quartzite, siliceous, thick-bedded, fine- to medium-grained, white, lavender; minor shale; some dark-gray dolomite interbedded near top; local conglomerate at base	
LOWER ORDOVICIAN		
Unnamed argillite 200+ (present in Hailey quad.)	*Ramshorn slate* 0–2000+(?) Slate, dark green to purplish, banded; basal conglomerate; subordinate quartzite	

TABLE 1.—*Continued*

Geosyncline of Central Idaho SW. of Lemhi Range, Fig. 1 (after Ross, 1934a; 1934b; 1937; 1947, except where credited otherwise)		Cratonic shelf of SW. Montana NE. of Tendoy Range, Fig. 1 (after Sloss and Moritz, 1951; Scholten *et al.*, 1955)
Western belt (mainly eugeosynclinal) through W. Bayhorse quad. and west, Fig. 1	*Eastern belt* (mainly miogeosynclinal) extending slightly W. of Lost River Range, Fig. 1	
	CAMBRIAN	
Garden Creek phyllite **and** ***Bayhorse dolomite*** Phyllite, dark gray to black, sericitic, soft, locally slightly calcareous. Dolomite, light creamy gray, thick-bedded, with black chert nodules. Sloss (personal communication) reports *Girvanella*, confirms Cambrian age. (Present locally in northern part of Bayhorse quadrangle)		***Pilgrim (Hasmark) formation*** 150–225 Dolomite, massive, saccharoidal, partly oölitic, light gray, tan; minor shale near base ***Park shale*** 75–150 Shale, micaceous, fissile, greenish gray, some reddish brown; minor dolomitic limestone, edgewise conglomerate ***Meagher dolomite*** 500–600 Dolomite, massive, thin-bedded near top, saccharoidal; shaly and glauconitic near top ***Wolsey shale*** 100–150 Shale, micaceous, fissile, grayish green, with calcareous nodules; thin lenses of fine-grained sandstone ***Flathead quartzite*** 100–150 Quartzite, medium-bedded, locally cross-bedded, medium- to coarse-grained, well rounded and sorted, scattered quartz pebbles, siliceous cement; conglomeratic near base; reddish brown

considered the Bayhorse the younger unit. Confirmation of the Cambrian age comes from Sloss (Personal communication), who reports the presence of *Girvanella* in the Bayhorse dolomite and states that structural relations show this to be the older unit. It is impossible to estimate the original thickness of Cambrian strata deposited in this area.

PLATE 1.—ANGULAR UNCONFORMITY BETWEEN PRECAMBRIAN AND ORDOVICIAN SEDIMENTARY ROCKS

FIGURE 1.—Exposure in cliff on north wall above mouth of Burnt Canyon, Beaverhead Range, Idaho. View looking west. Dipping rocks on left are purplish, argillaceous Belt sandstone. Higher, flat-lying rocks to the right are white to pink Kinnikinic quartzite, partly gravelly. Lemhi Range in background.

FIGURE 2.—Close-up view above Burnt Canyon, Beaverhead Range, Idaho. Head of right hammer marks angular unconformity. Head of left hammer rests on bedding plane in Belt sandstone.

FIGURE 1

FIGURE 2

ANGULAR UNCONFORMITY BETWEEN PRECAMBRIAN AND ORDOVICIAN SEDIMENTARY ROCKS

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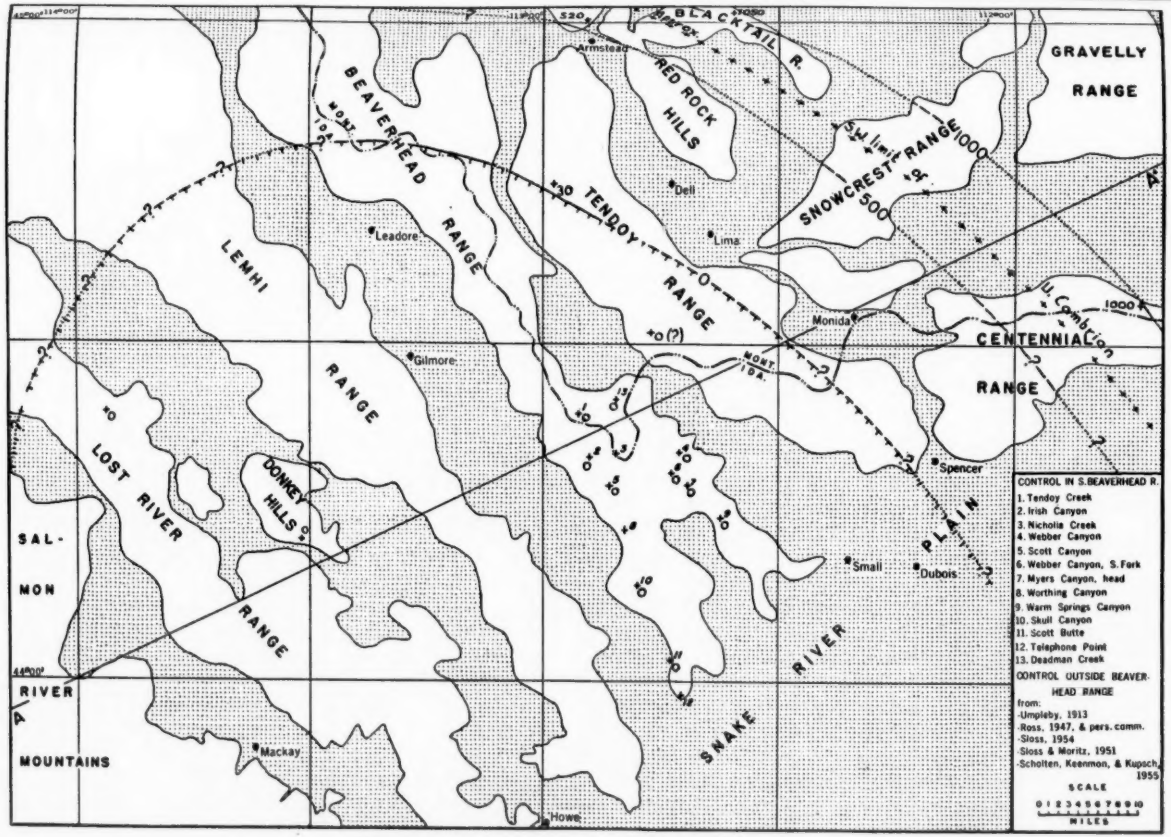

FIGURE 2.—ISOPACH MAP OF THE CAMBRIAN SYSTEM
AA′ is line of cross section in Figure 7.

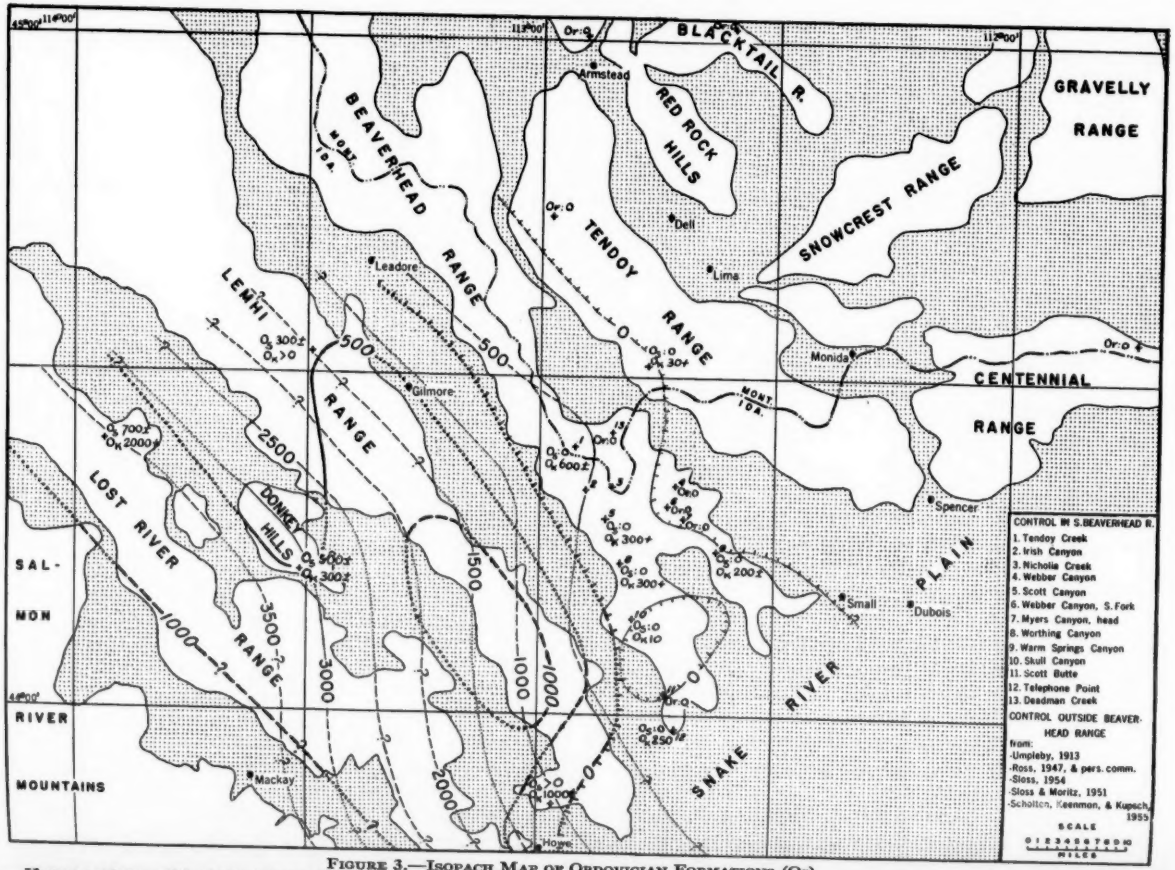

FIGURE 3.—ISOPACH MAP OF ORDOVICIAN FORMATIONS (Or)

Heavy contours with slanted figures represent Saturday Mountain dolomite (O_{sm}), other contours represent Kinnikinic quartzite (O_K).

Ordovician System.—In central Idaho west of the Lost River Range the Ordovician section is nearly complete and totals close to 10,000 feet. The Lower Ordovician Ramshorn slate has been mapped west of the Lost River Range but probably does not occur in the range itself (Ross, 1934a, p. 945–946; personal communication). No rocks of that age exist in the southern Beaverhead and Tendoy ranges or anywhere in southwestern Montana.

The Middle to Upper Ordovician Kinnikinic quartzite thins eastward from several thousand feet in the Lost River Range to about 1000 feet in the southern Lemhi Range (Sloss, 1954, p. 366; Ross, personal communication) and wedges out along a sinuous line through the Beaverhead and Tendoy ranges (Fig. 3). It is also absent from a small, isolated area along the west slope of the southern Beaverhead Range.

The Saturday Mountain dolomite (Upper Ordovician) thins from 3000 to 500–700 feet eastward into the Lost River Range. It is several hundred to more than 1000 feet thick in the southern Lemhi Range, thickens northward (Fig. 3; Ross, 1934a, Fig. 2; 1947, p. 1105; personal communication), but thins again to about 500 feet northwest of Gilmore according to a reconnaissance observation by Umpleby (1913, p. 33). Sloss (1954, p. 366–368) states that the same formation, to which he applies the southeast Idaho name Fish Haven, is present on the west flank of the southern Beaverhead Range. The writer has examined numerous sections in this area and in many places has observed several tens of feet of dark dolomite between the Kinnikinic quartzite and Mississippian rocks, which could belong to the Saturday Mountain. However, they are more similar to the Devonian Jefferson formation, and Late Devonian fossils occur only 60 feet above the Kinnikinic. Hence little, if any, Saturday Mountain dolomite appears to be present in the southern Beaverhead Range. Apparently the formation wedges out rather abruptly eastward and southeastward. It is not present in southwestern Montana.

Silurian System.—In the eugeosynclinal belt west of the Lost River Range the clastic rocks of the Trail Creek formation represent early Middle Silurian time, but near and in the Lost River Range only the younger Middle Silurian Laketown dolomite appears to be present, with observed thicknesses of 325–3000+ feet (Ross, 1934a, p. 956–960; 1947, p. 1106). Umpleby (1913, p. 33) reports about 200 feet of Laketown on the east flank of the Lemhi Range near Meadow Lake, northwest of Gilmore, but the formation seems to be absent in the southern part of that range (Ross, personal communication; Sloss, 1954, p. 366–368). Thus, regional eastward thinning is indicated (Fig. 4).

The Laketown dolomite probably is absent in the southern Beaverhead Range also, although a few isolated exposures of light-colored dolomite overlying the Kinnikinic in fault blocks above Worthing Canyon may be Laketown. As none occurs elsewhere in the southern Beaverhead Range or farther east it seems more likely that this light dolomite belongs to the Jefferson or Grand View formation (Devonian).

Devonian System.—Middle to Upper Devonian rocks occur throughout the miogeosynclinal belt of central Idaho, where they are represented by the Jefferson and Grand View dolomites. In addition, the overlying Three Forks limestone, which varies erratically in thickness and lithology but is mostly less than 300 feet, has been recognized in the Lost River Range, though not farther west. The Devonian strata thin eastward from 3000–3500 feet in the Lost River Range to a few hundred feet in southwestern Montana (Ross, 1947, p. 1108–1111; Sloss and Moritz, 1951, Fig. 8). They are several hundred to more than 1000 feet thick in the southern Lemhi Range, thickening northward (Ross, personal communication).

On the west flank of the southern Beaverhead Range are medium- to dark-gray and rusty dolomites which resemble the Jefferson. In some places fossils of Late Devonian age occur in an overlying unit of thin-bedded limestone, evidently the Three Forks formation. The writer collected such a fauna in Skull Canyon about 60 feet above the Kinnikinic quartzite. F. M. Swartz (Personal communication) identified *Camaratoechia*? sp., *Pugnoides* sp., and *Tenticospirifer* aff. *utahensis*. Shenon (1928, p. 7) reports Late Devonian fossils in Worthing Canyon. Three Forks and Jefferson aggregate 100–300 feet in most places. At the southern tip of the Beaverhead Range 331 feet, mostly

FIGURE 3.—ISOPACH MAP OF ORDOVICIAN FORMATIONS (Or)
Heavy contours with closed figures represent Saturday Mountain dolomite (Os), other contours represent Kinnikinic quartzite (Ok).

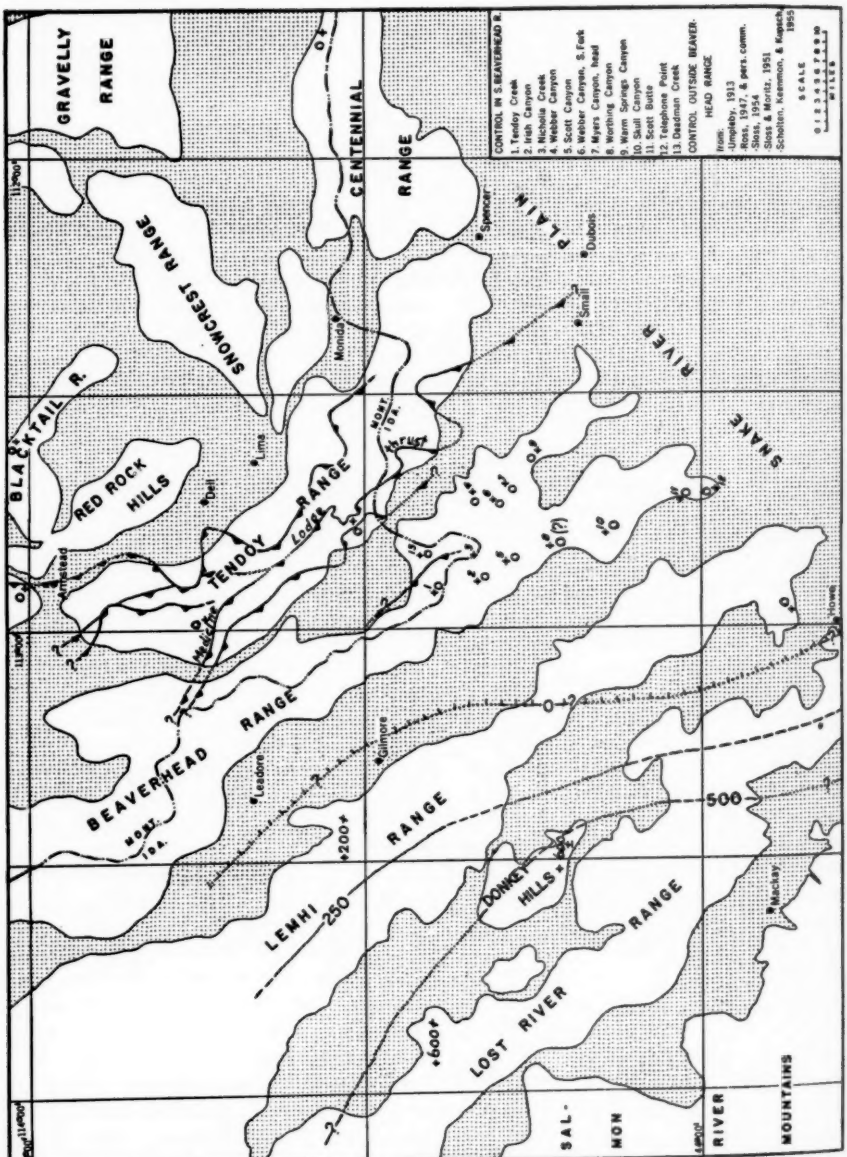

FIGURE 4.—ISOPACH MAP OF THE SILURIAN SYSTEM

Also shown are Laramide overthrusts in the Tendoy and southern Beaverhead ranges.

FIGURE 4.—ISOPACH MAP OF THE SILURIAN SYSTEM
Also shown are Laramide overthrusts in the Tendoy and southern Beaverhead ranges.

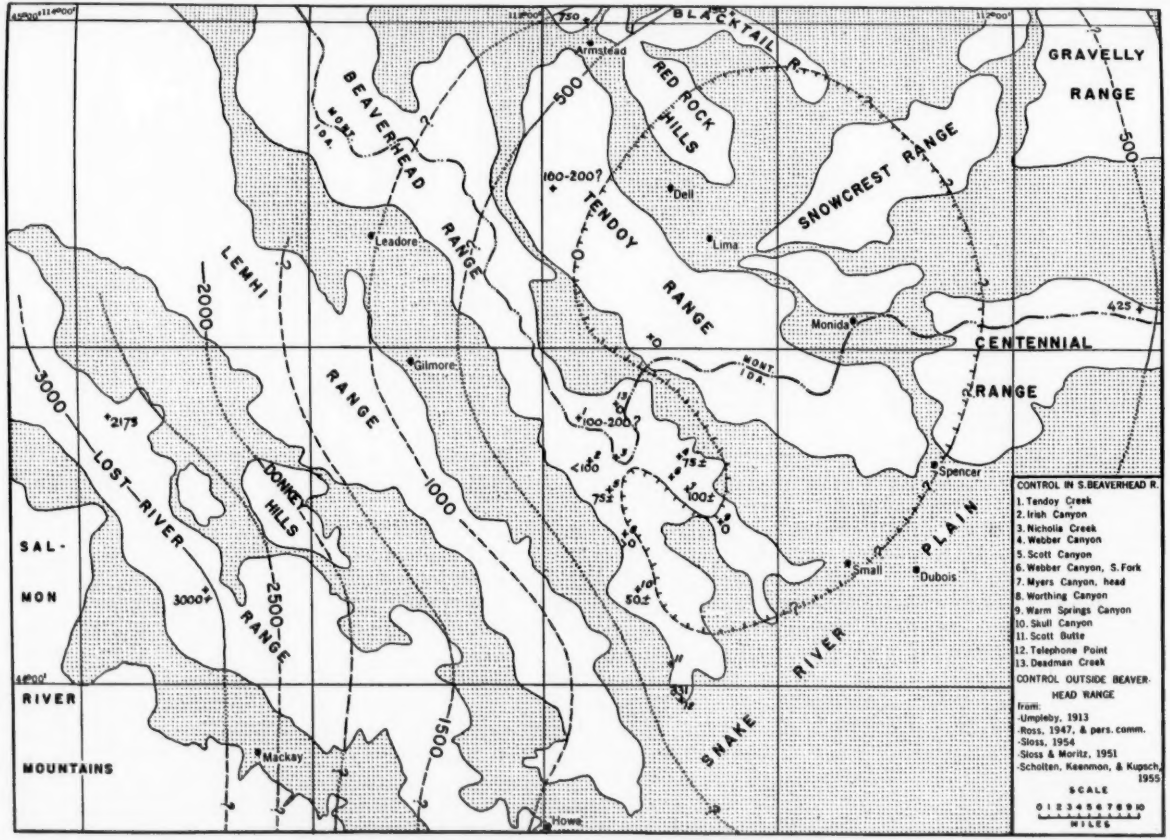

FIGURE 5.—ISOPACH MAP OF PRE-THREE FORKS DEVONIAN STRATA
In the southern Beaverhead Range some Three Forks beds may locally be included.

Jefferson-type dolomite, was measured between the Kinnikinic and the Mississippian strata. The Devonian wedges out near Warm Springs Canyon (Fig. 5). Across the border in Montana typical Three Forks occurs in poor exposures or float, but Jefferson is absent over a large area. It reappears farther north and east in southwestern Montana.

Mississippian System.—The Mississippian strata in the region between central Idaho and southwestern Montana present difficult problems of detailed correlation. In southwestern Montana the Lower Mississippian is represented by the Madison limestone, which may include beds of Meramec age and is bounded at the top by an erosional unconformity (Sloss and Moritz, 1951, p. 2157). The Madison occurs as far west as the Tendoy Range and has been provisionally mapped in a single small area in the southern Beaverhead Range (Scholten *et al.*, 1955, Pl. 1). However, the writer has since found post-Madison fossils in many limestone exposures in the southern Beaverhead Range which closely resemble the Madison, and he now inclines toward the belief that Madison is confined to the area east of the southern Beaverhead Range, where its thickness ranges from 1000 to 2000 feet.

The dark shales and shaly limestones of the Milligen formation in central Idaho are roughly correlative with the Madison (Ross, 1937, p. 32–33; 1947, p. 1113), although Weller *et al.* (1948, p. 140) list a possible age range from latest Devonian to middle Chesterian. According to Sloss (Personal communication) the upper Milligen west of the Lost River Range is probably Pennsylvanian. The age, correlation, distribution, and thickness of this unit in the mountains to the east are all points at issue. Ross (1947, p. 1113) states that the Milligen thins eastward from 3000 feet in the Bayhorse region to 1000 feet in the Lost River Range. Sloss (1954, p. 366) ascribes to the Milligen an incomplete section of 200 feet of black, siliceous argillite overlying Devonian beds with *Cyrtospirifer* near the southern tip of the Lemhi Range. Ross (Personal communication) assigns these to the younger Brazer formation and questions whether the name Milligen can be attached to any rocks in the Lemhi Range.

In the Beaverhead Range C. A. Moritz, W. O. Kupsch, and the writer measured more than 2000 feet of dark shale and shaly limestone at the head of Nicholia Creek along the Montana State line and applied the term Milligen to the unit (Sloss and Moritz, 1951, p. 2161; Scholten *et al.*, 1955, Pl. 1). In many other places in the southern Beaverhead Range of Idaho the writer has observed a stratigraphic interval composed of siliceous, mostly dark argillites, shaly limestone, some shale with worm tracks, and, locally, some coaly shale, below a thick series of cherty limestones. The interval is commonly up to several hundred feet thick but wedges out near Warm Springs Canyon (Fig. 6).

For the purpose of the present paper, the question of the name for the argillaceous rocks above the Devonian in the Lemhi and Beaverhead ranges is less important than the question of their age and correlation—specifically, their relation to the Madison limestone farther east. According to Sloss (Personal communication), the top of the Milligen in central Idaho shifts downward toward the east, and in the Lemhi Range the base of the overlying Brazer limestone may range in age from a possible Osagean, through Meramecian, to Chesterian, as suggested by coral data. On the other hand, fossils collected by the writer from the lower part of the Brazer limestone in the southern Beaverhead Range appear to be latest Mississippian to Early Pennsylvanian (G. A. Cooper, personal communication). The writer therefore doubts that the Brazer in this region includes a significant thickness of strata older than Chesterian or late Meramecian. For the underlying argillaceous interval in the two ranges there now appear to be the following possibilities: (1) it is entirely late Meramecian to early Chesterian, hence post-Madison; (2) it is entirely pre-late Meramecian, corresponding to all or part of the Madison limestone farther east; (3) it includes rocks of both age ranges.

Because of this uncertainty, no unequivocal isopach pattern can be drawn for the Lower to Middle Mississippian (pre-late Meramecian) rocks in the Beaverhead and Lemhi ranges. In Figure 6 the two extremes implied by alternatives (1) and (2) are shown. The zero-isopachs labeled AA′ are drawn on the basis of assumption (1), whereby there would be a large

FIGURE 6.—ISOPACH MAP OF LOWER TO MIDDLE MISSISSIPPIAN STRATA

Milligen shale west of Lemhi Range, Madison limestone east of southern Beaverhead Range. Several hundred feet of argillaceous rock in these ranges may be Lower to Middle Mississippian (alternative (2), corresponding to zero isopach BB'), or Upper Mississippian (alternative (1), corresponding to zero isopach AA').

area with no pre-upper Meramecian Mississippian rocks in the southern Beaverhead and Lemhi ranges. If assumption (2) be correct, there would still be a small area in the southern Beaverhead Range (outlined by the alternative zero-isopach labeled BB′) where pre-upper Meramecian rocks do not occur. This area includes Warm Springs Canyon, where, as stated before, the argillaceous section wedges out. Here, limestones with up to 50 per cent chert in parallel bands rest directly on Beltian and Kinnikinic sandstones, in the absence of Devonian strata. Associated quartzitic and dolomitic rocks carry Late Mississippian to Early Pennsylvanian fossils. A similar unusual, but significant relationship had previously been observed in the vicinity of Deadman Creek, about 6½ miles north of the southermost tip of Montana (Scholten *et al.*, 1955, Pl. 1). At that time the absence of pre-Brazer Paleozoic strata had been attributed to thrust faulting, but it is now thought likely that this is a stratigraphic contact similar to that near Warm Springs Canyon. According to assumption (2), rocks older than upper Meramecian would be present outside the Warm Springs Canyon-Deadman Creek area but in most places would be comparatively thin (several hundred feet). The highly anomalous thickness of 2200 feet at Nicholia Creek may, perhaps, be ascribed to tectonic flowage during the Laramide orogeny, to local faulting, or to rapid interfingering with the lower Brazer limestone, rather than to variation in the amount of sediment deposited during a given time interval. Facies changes of this type have been observed by Ross (1937, p. 31) in the Brazer of central Idaho, and Sloss (Personal communication) describes the contact between the Milligen and Brazer as intertonguing.

Whichever correlation is correct, the total amount of sediment representing the Early to Middle Mississippian in the southern Beaverhead and Lemhi ranges appears to be, at best, surprisingly small in most places, as compared with the regions to the east and west, and in at least one area actually zero. Possibly this time span is not represented by any sediments in these ranges.

The Late Mississippian in much of central Idaho is represented by Brazer limestone. No complete exposures of this unit have been found, but Ross (1947, p. 1116; personal communication) has observed almost 4000 feet of Brazer in the Lost River Range and an estimated 2000–3000 feet in the Lemhi Range. Probably the true thickness is much greater. Westward, the Brazer wedges out into the eugeosynclinal belt. To the east, several thousand feet of Brazer is doubtlessly present in the southern Beaverhead Range, where the upper part has been removed by erosion. Lithologically, the formation, with its dark cherty limestones and interbedded quartzitic, dolomitic, and shaly units, resembles the type Brazer of southeastern Idaho, which appears to range in age from middle Meramecian to late Chesterian (Williams, 1943, p. 610; Mansfield, 1927, p. 63–71). However, part, if not most, of the Brazer of east-central Idaho may be post-Chester in age. Most of the brachiopods collected by the author are suggestive of an Early Pennsylvanian age according to G. A. Cooper (Personal communication), who tentatively identified *Spirifer* cf. *rockymontanus*, *Spirifer* cf. *composita*, *Spirifer* cf. *boonensis*, *Productus* cf. *arkansanus*, and a species of *Dictyoclostus* which probably belongs to the Early Pennsylvanian. Pennsylvanian, or even Permian, affinities among Brazer fossils had previously been noted by Ross (1934a, p. 992–993; 1947, p. 1116–1117) and Scholten *et al.* (1955, p. 365).

In southwestern Montana the Big Snowy group and overlying lower Amsden formation are Late Mississippian (mostly or entirely Chester) and range in combined thickness between 600 and 1000 feet, thinning eastward to zero (Sloss and Moritz, 1951, Fig. 10). Hence, rocks of Chester age show gradual thinning and comparatively rapid facies change eastward across the geosynclinal margin. Control is lacking to detect possible local deviations from this pattern.

Pennsylvanian System.—A Lower Pennsylvanian facies change is in evidence between the thin upper Amsden formation of southwestern Montana and the correlative part of the Brazer limestone of southeast-central Idaho.

Overlying these units are the Wood River formation of central Idaho and the Quadrant formation of the Tendoy Range and east, both dominantly sandstone and subordinately carbonate. The Wood River graywackes have been estimated at 8000 feet and have been mapped

as far east as the Lost River Range (Ross, 1934a, p. 988; 1947, p. 1117, Pl. 1). Bostwick (1955, p. 942) measured a possible maximum of about 12,000 feet in the Hailey quadrangle. The Quadrant quartzite attains approximately 2600 feet in the Tendoy Range and within 50 miles thins eastward to a few hundred feet (Sloss and Moritz, 1951, Fig. 11). It appears to be present in the Beaverhead Range (Scholten *et al.*, 1955, Pl. 1; Cressman, 1954, p. 191).

Fusulinids in the Wood River of the Hailey quadrangle indicate that the formation includes most of the post-Morrow or post-Atoka Pennsylvanian (Bostwick, 1955, p. 941, 947), which makes it the approximate correlative of the Quadrant formation (Sloss and Moritz, 1951, p. 2165). Possibly at least part of the Wood River of central Idaho and the Quadrant of southwestern Montana correlate in time with the upper Brazer in the intervening Lost River, Lemhi, and Beaverhead ranges, as suggested by the fossil data in the Brazer.

Permian System.—Early Permian (Wolfcampian) time is represented by the upper two-thirds of the Wood River formation of central Idaho as shown by fusulinids, and the Quadrant quartzite of southwestern Montana may include beds of this age also (Bostwick, 1955, p. 944; Sloss and Moritz, 1951, p. 2165). In the intervening area even the Brazer carries fossils which could be Permian (Ross, 1934a, p. 993). Blackstone (1954, p. 923–925) reports Wolfcampian fusulinids from limestones at the southern tip of the Lemhi Range; these rocks seem to resemble the Brazer more than any other unit, but a definite formation assignment must await more-detailed study.

The Phosphoria formation, which, in the type area of southeastern Idaho, probably is post-Wolfcampian (Cheney and McKelvey, 1956, p. 1717), is about 800 feet thick in the Tendoy Range and thins eastward (Scholten *et al.*, 1955, p. 366). Cressman (1954, p. 191) describes a single occurrence, 850 feet thick, on the west slope of the Beaverhead Range, east of Leadore, Idaho. In the Casto quadrangle of central Idaho (Fig. 1) several thousand feet of Casto volcanics is exposed; it is believed to be Permian, though this is not conclusively established (Ross, 1934a, p. 994–996; 1934b, p. 28–35). Blackstone (1954, p. 924–925) summarizes correlations made by others which would imply that the Casto volcanic rocks are of approximately the same age as the Phosphoria formation. The nearest known exposures of these two units lie about 60 miles apart. Assuming the correlation is correct, this fact, and the westward thickening of the Phosphoria into the Beaverhead Range, suggests that post-Wolfcamp Permian rocks may once have been continuous across east-central Idaho between the present areas of distribution of Phosphoria sediments and of Casto volcanic rocks.

Tectonic Evolution

Precambrian Background

During the late Precambrian the Beltian geosyncline included the area of the southern Beaverhead Range, where considerably more than 1000 feet of Belt sandstones and argillites was deposited. Evidently a well-defined geosynclinal hinge existed not far east of the Beaverhead Range. Sloss (1950, p. 432) believes an orogenic trend existed during Beltian time along the north edge of the Wyoming shelf, which would include southwestern Montana.

Early Paleozoic Development

During Middle and Late Cambrian time southwestern Montana was an area of slow subsidence on the edge of the cratonic shelf. Lack of data makes accurate reconstruction of Cambrian basin conditions in central Idaho impossible, but Sloss (1950, Fig. 4) represents them in a general way as being widespread and continuous.

The absence of Upper Cambrian, or Upper and Middle Cambrian strata over southeast-central Idaho and immediately adjacent parts of Montana (Fig. 2) has been ascribed primarily to pre-Middle Ordovician (pre-Kinnikinic) and pre-Middle Devonian uplift and erosion, but the local presence of topographically positive elements during the Cambrian has also been mentioned as a cause of nondeposition or erosion (Sloss and Moritz, 1951, p. 2139–2140; Sloss, 1950, p. 433–434). The angular unconformity between Belt and Kinnikinic strata in the southern Beaverhead and Lemhi ranges is evidence of a pre-Kinnikinic uplift which was especially sharp in this local area, representing one of the strongest Early

Paleozoic disturbances in the northern Rocky Mountain geosyncline. The terms "Skull Canyon Disturbance" and "Skull Canyon Uplift" are here proposed from the exposures of the unconformity in and near Skull Canyon in the southern Beaverhead Range. As the Belt strata in this vicinity are tilted northward and northeastward and the overlying Kinnikinic is almost horizontal, the center of the uplift may have been south to southwest of Skull Canyon. Sloss (Personal communication) states that a local downwarp in the Garden Creek area of the Bayhorse quadrangle was responsible for preservation of Cambrian and Lower Ordovician (Ramshorn) strata below a pre-Middle Ordovician erosion surface. Presumably this warping was also part of the Skull Canyon disturbance, which resulted regionally in lifting a broad area in Idaho and Montana above sea level.

Subsidence in southeast-central Idaho was resumed during Middle and Late Ordovician time. Depression of geosynclinal magnitude occurred as far east as the Lemhi and, perhaps, the west slope of the Beaverhead Range, but any Ordovician strata farther east were undoubtedly thin, and southwestern Montana remained essentially stable. In the southern Beaverhead Range the local occurrence of Belt cobbles in coarsely sandy to gravelly basal Kinnikinic near Skull Canyon indicates that a topographic high, a last remaining testimony to the Skull Canyon disturbance, was still in existence near by, probably to the south. The thinning to zero of Kinnikinic toward an isolated area south of Skull Canyon could be at least partly ascribed to a persistence of this high through Kinnikinic time. However, there is some evidence of post-Middle Silurian erosion in this region, and it may therefore be that upper Kinnikinic finally overlapped across this local high and was removed later.

From latest Ordovician through the Silurian the southern Beaverhead region appears to have been relatively stable on the edge of the cratonic shelf, receiving at most thin covers of Saturday Mountain and Laketown dolomite. Any sediments deposited during this period were removed or almost removed before the Late Devonian. In the Donkey Hills and adjacent Lemhi Range the isopach pattern of the Saturday Mountain (Fig. 3) suggests that an isolated arch may have existed in this area before deposition of the Laketown, which does not reflect the pattern. However, data are too meager to make this a reliable conclusion.

Mid-Paleozoic Development

Sloss (1954, p. 368) has postulated the rise of a pre-Late to Late Devonian feature, termed "Lemhi Arch," at the edge of the geosyncline, to explain the rapid thinning of the Saturday Mountain (or Fish Haven) and Laketown dolomites into the southernmost part of the Lemhi Range, and the alleged occurrence of thicker sections of these formations farther east. However, the writer has found that both units are very thin or, more likely, absent in the southern Beaverhead Range, and he believes evidence for an isolated arch in this part of Idaho is lacking. The isopach patterns of the Saturday Mountain and Laketown dolomites do not suggest greatly anomalous tectonic conditions in the southern Lemhi Range and vicinity (Figs. 3, 4). Rather, they are those that would be expected to result from deposition at the edge of a geosyncline, with nondeposition or subsequent erosion on an emergent cratonic shelf in the area of the Beaverhead Range and east.

A suggestion of post-Laketown erosion is afforded by the rapid disappearance of the Saturday Mountain dolomite between the southern Lemhi and Beaverhead ranges, and of the Laketown dolomite between the Lost River and southern Lemhi ranges. Over most of the region any such erosion must have taken place in pre-Middle Devonian (pre-Jefferson) time, but, in the general vicinity of the southern Tendoy Range, emergence evidently continued into Middle and Late Devonian time. This would account for the absence of pre-Three Forks Devonian strata by nondeposition in this vicinity (Fig. 5). The alternative possibility, Devonian deposition followed by pre-Three Forks erosion, seems less likely in view of the absence of any evidence of erosion in the immediately surrounding area, where Three Forks rests on Jefferson with apparent conformity (Sloss and Moritz, 1951, p. 2152; Scholten *et al.*, 1955, p. 363). In either case, the evidence indicates the existence of a Devonian upwarp in this region. The inferred position of the zero

isopach of the pre-Three Forks Devonian (Fig. 5) suggests a connected high in the Skull Canyon area, but correlations are too unreliable to confirm this.

From this it appears that the arch postulated by Sloss is not completely erased as a Paleozoic tectonic element, but it is located farther east, in a closer relationship to the cratonic shelf than had been inferred, and it extended through Jefferson time. If the upwarp had a pre-Jefferson history as a tectonic entity, widespread post-Laketown erosion on the shelf has removed all evidence for this. Perhaps the term Tendoy Dome would be more appropriate for the upwarp, in view of its geographic relation to the Tendoy Range and its apparent subcircular outline.

Jefferson strata were deposited around this upwarp, which was finally overlapped by Three Forks sediments in latest Devonian time. Some sandy and gravelly units with pebbles of Kinnikinic quartzite occur below Jefferson dolomite in the southern Beaverhead Range. The age of these units is uncertain, but they may be basal Jefferson and represent a testimony to Devonian erosion on the dome.

During the Early to Middle Mississippian the cratonic shelf in southwestern Montana was subject to greatly increased subsidence with deposition of thick Madison limestone. In the southwest corner of the State, depression reached geosynclinal magnitude. Still farther west the Madison basin merged into the eugeosynclinal Milligen basin of east-central Idaho (Fig. 6). It is necessary, however, to postulate positive movements during or shortly after this time to account for the rapid thinning or disappearance of this time-rock unit in the southern parts of the Beaverhead and Lemhi ranges. This high probably took the form of a northwest-trending arch, but its detailed extent and outline are in doubt because of difficulties of correlation in this area. The two alternative courses of the zero-isopach contour sketched in Figure 6 outline two possible positions for the highest part of the uplift, which is here referred to as the Beaverhead Arch. This postulated arch roughly separated the area of Milligen from that of Madison deposition. Possibly it was not fully developed in the earliest Mississippian (Kinderhookian), thus allowing fine clastics to reach the Madison basin and become incorporated in the shaly Lodgepole (lower Madison) limestone of the Tendoy Range and east. Such a continuity has been suggested by Sloss and Moritz (1951, p. 2161). Post-Kinderhook rise of the arch may then have effectively sealed off the Madison basin from the further influx of clastics from the west, permitting deposition of the much-purer, thick Mission Canyon (upper Madison) limestone at the same time that great volumes of argillaceous sediment, derived from a westerly source, accumulated in east-central Idaho.

In late(?) Meramecian time widespread uplift, weathering, and erosion occurred in the region of Madison deposition as far west as the Tendoy Range (Sloss and Moritz, 1951, p. 2139–2140; Scholten *et al.*, 1955, p. 365), before deposition of the variable shelf-margin strata of the Big Snowy group and lower Amsden formation during the Chesterian epoch. In central Idaho geosynclinal subsidence persisted uninterruptedly into and through Chesterian time. Deposition of Milligen(?) argillites may have continued into early Chester time and overlapped the Beaverhead arch to connect with the Big Snowy basin to the east. Sloss and Moritz (1951, p. 2161) believe the Big Snowy shales may represent an eastern extension of the Milligen environment of sedimentation.

If this chronology of the Beaverhead arch is correct, the post-Devonian argillaceous interval in the southern Lemhi and Beaverhead ranges could be partly Kinderhook and partly early Chester in age, with nondeposition during Osagean and Meramecian, when the arch was above effective wave base for fine clastics. In the Warm Springs Canyon-Deadman Creek area, where there are no argillaceous beds, the arch may have risen above wave base somewhat earlier and persisted somewhat later. However, before the close of the Mississippian thick Brazer limestones and quartzitic beds were accumulating across the entire Beaverhead Range area.

Late Paleozoic Development

The thick Chester and Pennsylvanian strata in this part of the northern Rocky Mountains have been insufficiently studied, subdivided, and correlated to allow precise conclusions concerning the paleotectonic patterns during this time. In latest Mississippian to earliest Pennsyl-

vanian time the shelf still extended as far west as the Tendoy Range, where the Amsden strata were laid down while Brazer sediments accumulated in the geosyncline to the west. Soon thereafter, the tectonic hinge migrated eastward, and the Tendoy Range area was geosynclinally depressed with deposition of the Quadrant quartzites, while monotonous sedimentation of Wood River graywackes took place in central Idaho. Possibly, Brazer limestones and quartzitic deposits continued to accumulate in the intervening region of the Beaverhead, Lemhi, and Lost River ranges.

During the Early Permian (Wolfcampian) geosynclinal subsidence continued in the Wood River basin in central Idaho and, probably, in the area of Quadrant deposition in extreme southwestern Montana. Little can be said with confidence concerning tectonic conditions during the post-Wolfcamp Permian. The thickness of the Phosphoria formation nowhere attains the order of magnitude usually associated with a geosynclinal environment, but this may be the result not so much of a decreasing rate of subsidence as of slower deposition in a relatively deep sea. McKelvey *et al.* (1953, p. 57, 59, Fig. 1) believe phosphorite accumulated in the absence of a large influx of clastics in a water depth of 600 to 3000 feet, and they state that the approximate amount of P_2O_5 in the Phosphoria would be contained in about 5000 feet of sediment with a normal percentage of P_2O_5. They depict the Phosphoria of extreme southwestern Montana as lying in a miogeosynclinal zone with a hinge passing east of the Tendoy Range. The position of the western margin of this Permian basin is unknown. The existence of several thousand feet of volcanic rocks in the Casto quadrangle does not imply geosynclinal downwarping, and the edge of the basin probably lay to the east.

Laramide Sequel

The outstanding structural characteristic imposed on this region by the Laramide orogeny is a prominent zone of major overthrusts trending northwest through the Tendoy Range (Fig. 4; Scholten *et al.*, 1955, Pls. 1, 4, 6). The most prominent is the Medicine Lodge overthrust, a low-angle fault with a displacement of many miles and traceable over more than 40 miles. This belt appears to be a continuation of the thrust belt of southeastern Idaho, and it has been suggested that the Medicine Lodge overthrust may be a northern extension of the Bannock overthrust (Kirkham, 1927, p. 27). Thrusting occurred also to the west in the region of the Paleozoic geosyncline of central Idaho; thrust faults have been mapped in parts of the Beaverhead and Lost River ranges (Scholten *et al.*, 1955, Pl. 1; Ross, 1947, Pl. 1). They do not, however, appear to have a continuity or displacement comparable to the thrusts of the Tendoy Range. In the area of the Paleozoic shelf adjoining the Tendoy Range on the east, a wide zone with only one, northeast-trending, thrust (Klepper, 1951, Pl. 16) separates the Tendoy thrust belt from the thrust region north and northwest of Yellowstone Park.

Hence, it appears that the Paleozoic hinge zone had a reaction to the Laramide orogeny different from adjacent geosynclinal and shelf regions. In this zone there was a greater tendency for compressive stresses to be taken up along faults, rather than by folding with only minor thrusting.

Summary and Conclusions

Facies variations, over-all thickness changes, and unconformities in the Paleozoic of southwestern Montana and southeast-central Idaho reveal the existence of a geosynclinal basin bordered on the east by a semistable cratonic shelf, with a hinge zone which fluctuated approximately between the areas of the present Lemhi and Tendoy ranges, tending to migrate eastward during the Carboniferous.

Whereas this pattern prevailed during most of the Paleozoic era, it is now apparent that the area of the southern Beaverhead and Tendoy ranges, at times, did not conform to it. Although later work will undoubtedly necessitate changes in the tentative isopach patterns presented in Figures 2–6, it is, nevertheless, clear that this area was one of recurrent uplift during the Paleozoic. The Skull Canyon Uplift existed prior to the Middle Ordovician, with a center of uplift possibly somewhere south or southwest of Skull Canyon on the west slope of the Beaverhead Range. A broad upwarp (Tendoy Dome)

FIGURE 7.—STRATIGRAPHIC CROSS SECTION OF PRE-UPPER MISSISSIPPIAN ROCKS
Line of cross section is shown on Figure 2.

seems to have come into existence in and around the southern Tendoy Range during Devonian time, remaining active up to the latest Devonian, when it was finally overlapped. Renewed uplift in the Early to Middle Mississippian raised the Beaverhead Arch in the area of the southern Beaverhead and Lemhi ranges. The composite stratigraphic relationships resulting from these disturbances are illustrated in the stratigraphic cross section (Fig. 7).

The causes for these uplifts are not clear. This problem is further explored in another paper (Scholten, 1956). It suffices to state here that the fact that these uplifts took place in the general zone of fluctuation of the tectonic hinge between geosyncline and shelf may be of genetic significance. The hinge could be considered a zone of weakness in the crust which was especially responsive to early stresses in the geosynclinal region. The development, during the Laramide orogeny, of a prominent zone of eastward thrusting approximately along the east margin of the area of Paleozoic upwarping also may be more than fortuitous. The Paleozoic disturbances may well have predestined the stronger diastrophic events of the Early Tertiary.

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PRELIMINARY STUDY ON THE REGIMEN AND MOVEMENT OF THE TAKU GLACIER, ALASKA

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ABSTRACT

The mathematical theory of equilibrium glaciers is applied to data obtained on the Taku Glacier in the Juneau Ice Field. Although the data are not detailed enough to accurately characterize the behavior of the glacier, they are sufficient to illustrate the methods used to obtain relations between accumulation, ablation, and flow of ice.

Measurements made on the net accumulation and distribution of snow at the end of the 1953 ablation season show that the quantity of snow was primarily a function of the elevation above the firn line.

Ablation of bare ice below the firn line was determined. About 40 feet of ice melts each year in the terminal area of the Taku.

Detailed studies of movement were made on three transverse profiles where the thickness of the ice had been measured by seismic methods. The volumes of ice passing through these profiles were calculated from three types of measurements—flow, ablation, and accumulation. The volumes calculated from the accumulation data are much larger than the volumes determined from either ablation or flow measurements. The volumes derived from ablation data are larger than those determined from movement data except near the terminus.

The Taku Glacier may be in an approximately equilibrium condition in the terminal region, but much more snow is accumulating in the upper regions of the glacier than is melting below the firn line.

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INTRODUCTION

Published data on most glaciers are insufficient to allow an accurate and detailed mathematical treatment of their regimen and flow. Enough information has been obtained about the Taku Glacier between 1948 and 1955 to allow approximate calculations and to illustrate how the methods of mathematics may be applied to study the behavior of a glacier. Even in this case, however, the data are inadequate for a complete mathematical analysis of regimen and flow.

One main objective of the 1953 Juneau Ice Field Research Project was to collect the data necessary to determine whether or not the Taku is in a state of equilibrium. At equilibrium the excess of snow for an average budget year

in the accumulation area is exactly equal to the average amount of bare ice that melts each year below the firn line. Glacier flow can transport the same amount of ice from the accumulation area into the ablation area to maintain the balance. The mathematical theory of equilibrium glaciers has been discussed in a previous paper (Nielsen, 1955). To apply the theory it is necessary to know the dimensions of the glacier, the amount and distribution of the snow cover remaining at the end of a typical ablation season, the quantity of bare ice melting each year below the firn line and its variations along the glacier, and detailed data on movement at several locations. Additional information to verify the internal congruity of the results can be obtained by using the surface slopes of the glacier. In demonstration of the general theory of an equilibrium glacier, the average values for the snow accumulation and ablation of ice, measured over a period of years, should be used rather than the values for a given year. Most of the data presented in this paper apply only to the year 1953, but data available from previous years were also used. The records of the Juneau Ice Field Research Project indicate that 1953 was an "average" year on the Taku, therefore it is reasonable to assume that errors resulting from lack of information are not great enough to invalidate the conclusions. It is desirable, however, that additional measurements be made to verify the results given here.

Taku Glacier, northeast of Juneau, Alaska, and the largest glacier of the Juneau Ice Field, has a maximum length of 36 miles and covers a range of altitudes from tide water to over 6000 feet. About 225 square miles of the glacier are above and 45 square miles are below the average late summer firn line, which averages about 3400 feet in elevation. During the past 50 years the Taku has advanced 3½ miles, and is still advancing very slowly. A sketch map of the Taku is given in Figure 1.

Detailed studies of movement were made on three transverse sections on the Taku. Seismic determinations of depth were made on these same three sections by Poulter in 1949 (Poulter, 1949, p. 1–12). In Figure 1 the sections are shown to be about 2 miles, 11 miles, and 17 miles from the terminus and are designated as profiles *1*, *2*, and *3*, respectively. The first two profiles are below the firn line, and the third just above it.

Let V_i be the volume discharge of ice through the cross section i of the main branch of the Taku where i can be *f*, *1*, *2*, or *3* representing the firn line, profiles *1*, *2*, or *3*, respectively. The ablation or wastage per year between the terminus and section i is W_i. The average velocity of flow at section i is $\bar{v}_i$; C_i and $\bar{D}_i$ are the width of the glacier and its average depth across the same section. The net average yearly accumulation above section i is A_i.

For an equilibrium glacier, the following equations should hold:

$$V_1 = W_1 = \bar{v}_1 C_1 \bar{D}_1 \quad (1)$$
$$V_2 = W_2 = \bar{v}_2 C_2 \bar{D}_2 \quad (2)$$
$$V_3 = A_3 = \bar{v}_3 C_3 \bar{D}_3 \quad (3)$$

The west branch and the northeast ("Demorest") branch enter the main Taku between sections *2* and *3*. If V_k is the volume of ice entering the main Taku from these branches, then the following additional equations are approximately true:

$$V_1 = V_f + V_k + W_1 - W_f \quad (4)$$
$$V_2 = V_f + V_k + W_2 - W_f \quad (5)$$
$$V_f + V_k = W_f \quad (6)$$

These are simplified equations for an equilibrium glacier. They will not be discussed in detail as they have been described previously (Nielsen, 1955, p. 3–5).

Acknowledgments

Most of the data presented here were obtained with the assistance of the various members of the 1953 Juneau Ice Field Research Project. This project is directed by the American Geographical Society through contract with the Office of Naval Research. Special thanks are due Austin Post, the party's surveyor. Dr. Calvin Heusser, William O. Field, and others connected with the Project have been of great help. The assistance given by the Air Force 10th Air Rescue Group, the Department of Defense, and the U.S. Forest Service was very valuable. Professor Robert Sharp gave a number of helpful suggestions.

Velocity of Flow on the Taku

Detailed studies of movement were made at three locations shown in Figure 1. At these same locations Poulter (1949, p. 5–12) had made seismic determinations of the ice depth. Six to 10 tripods to determine movement were installed across the glacier on each profile. In addition, other tripods were placed half a mile below the first seismic profile, and a row of stakes was placed 1 mile below the second seismic profile. Two surveys of the tripods were made a month to 6 weeks apart with a Wild T2 theodolite. The dates of the survey were July 11 and August 22 for the first profile, June 26 and July 25 for the second profile, and July 1 (?) and August 2 for the third profile. A third survey of several of the tripods on the first profile was made on July 9, 1955—2 years after they were originally surveyed.

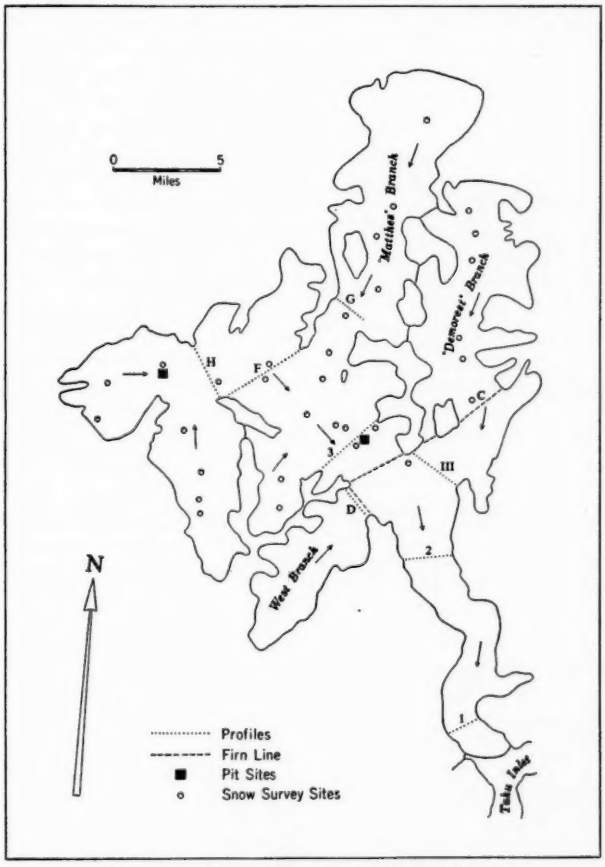

Figure 1.—Sketch Map of the Taku Glacier

Figures 2, 3, and 4 show some of the data on the horizontal component of the movement at the first, second, and third profiles, respectively. These results show the Taku to be very active. In Figure 4 at the third profile not only the results for 1953 are shown but also similar measurements made in 1952 (Gilkey, 1953, p. 14) and in 1950 (Miller, 1954, p. 43). The velocity of flow for 1953 appears to be somewhat less than for the other years; this

decrease may or may not exist because the date of the 1953 survey at this profile was omitted from the data book, and a constant percentage error may be involved. The errors in the velocity measurements on the first and second profiles, however, are less than 5 per cent.

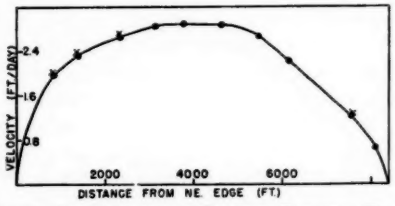

FIGURE 2.—VELOCITY OF SURFACE FLOW ON FIRST PROFILE, ABOUT 2 MILES FROM TERMINUS OF TAKU GLACIER

● 1953 data; × 1953–1955 data.

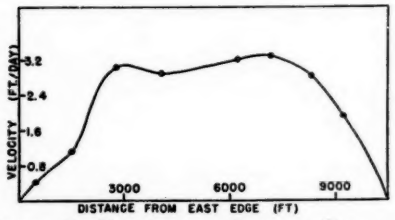

FIGURE 3.—VELOCITY OF SURFACE FLOW ON SECOND PROFILE, ABOUT 11 MILES FROM THE TERMINUS

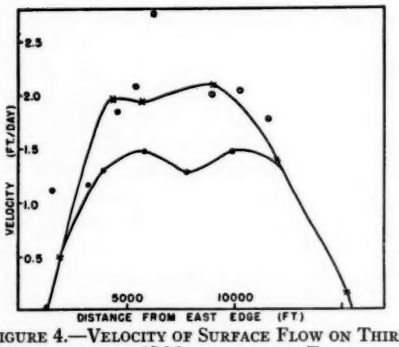

FIGURE 4.—VELOCITY OF SURFACE FLOW ON THIRD PROFILE, ABOUT 17 MILES FROM THE TERMINUS

● 1953 data; ○ 1950 data; × 1952 data.

The velocity of flow per day as measured during the summer of 1953 was assumed to be the same as the average daily flow throughout the whole of 1953 and also throughout the past few years. The successive positions, in both summer and winter, of an aluminum pipe sunk into the ice near the third profile since 1950 (Miller, 1954, p. 60) seem to indicate that the Taku moves at about the same rate throughout the year; the data obtained during the summer should be representative of the yearly average. Near the third profile the aluminum pipe moved about 1.91 feet per day between 1950 and 1952 and about 2.03 feet per day between 1952 and 1953. These values agree with the data shown in Figure 4 for the summer of 1953.

Additional evidence for a constant rate of flow throughout the year was obtained when some of the 1953 tripods on the first profile were resurveyed in July 1955. In Figure 2 the rate of flow during the 2-year period is compared with the rate during the summer of 1953. The two sets of data are nearly identical. The slightly greater values for the period 1953–1955 can be attributed to the fact that the Taku has increased in activity. In 1953 the Taku was almost stationary or actually retreating; in 1955 it was again slowly advancing along its edges, and the ice had thickened about 20 feet.

TRANSPORT OF ICE ALONG THE TAKU GLACIER

The quantity of ice transported through the cross sections at the profiles can be calculated from the data on movement, the seismic depth measurements of Poulter, and the measured widths of the glacier.

Since the velocity of flow changes across the glacier and with the depth, an average velocity of flow must be obtained by what is essentially a mathematical double integration. In this paper an average surface velocity was obtained by graphical integration. The average velocity throughout the glacier was then obtained by using the following equation derived by Nielsen (1955, p. 8).

$$\frac{\bar{v} - v_b}{\bar{v}_0 - v_b} = \frac{n+1}{n+2} \qquad (7)$$

In this equation $\bar{v}$ is the desired average velocity, $\bar{v}_0$ is the average surface velocity, and v_b is the velocity at the base of the glacier (slippage along the bottom). Quantity n is a constant dependent upon the properties of ice;

n is *1* for a viscous material. On the basis of Glen's experimental work (Glen, 1953, p. 721), $n = 3$ was used in this paper. It was assumed that v_b was 25 per cent of $\bar{v}_0$; v_b is an unknown quantity, but the desired velocity is very insensitive to the value of v_b. Table 1 shows the calculated average velocities, width of the glacier, and average thickness of the ice as found by Poulter. The product of the average thickness and the surface width of the glacier gives the cross-sectional area through which the ice is flowing. Only the width in which ice is actually moving, not the total width of the glacier, is used; the stagnant ice at the edges is neglected. Such a procedure is justified because Poulter's seismically computed thicknesses indicate that the Taku is many times wider than the depth; the width at the bottom is nearly the same as that on the surface. Thus, no corrections, other than those made in averaging the surface velocity are necessary. Therefore the quantity of ice transported through a cross section is:

$$V = C\bar{D}\bar{v} \qquad (8)$$

where V is the volume of ice passing through the cross section each year, C is the width of the glacier, $\bar{D}$ is the average thickness of the ice, and $\bar{v}$ is the average velocity of flow per year. The volumes V are listed in Table 2.

It would be desirable to have some data on the higher regions of the Taku and its branches, but very little information is available for the calculations of the volume of ice being transported except for some flow measurements by Argus (1952, data on file, American Geographical Society). Assuming that the average depth of the ice is 1000 feet, the results given in Table 3 were obtained from movement studies on the profiles which are shown in Figure 1 by the letters *C*, *D*, *F*, *G*, *H*. These results are very rough approximations, but they indicate that 16×10^9 cubic feet of ice per year should pass through the second profile and 12×10^9 cubic feet per year through the third profile. These values are considerably greater than the measured values as calculated from the velocity across the second and third profiles. Not enough data are available to explain this apparent discrepancy.

TABLE 1.—MOVEMENT STUDY PROFILES

Profile	Average bulk velocity $\bar{v}$ (ft./year)	Width of glacier (ft.)	Average seismic depth (ft.)
1	672	8400	1170
2	711	10500	780
3	450	14300	959

TABLE 2.—VOLUMES OF ICE TRANSPORTED ACROSS PROFILES

Profile	From flow data (cu. ft./yr.)	From ablation data (cu. ft./yr.)	From accumulation data (cu. ft./yr.)
1	6.61×10^9	4.05×10^9	25.6×10^9
2	5.82×10^9	13.80×10^9	35.3×10^9
3	6.17×10^9	14.77×10^9	27.5×10^9

TABLE 3.—TRANSPORT OF ICE ACROSS PROFILES FROM MOVEMENT DATA

Location	$\bar{v}$(ft./yr.)	Volume (cu. ft./yr.)
Fourth seismic profile (northeast branch)	238	4.8×10^9
Profile C (middle northeast branch)	237	2.8×10^9
Profile F (upper main Taku)	350	7.2×10^9
Profile H (upper main Taku)	490	7.0×10^9
Profile G (north branch)	340	4.5×10^9
Profile D (west branch)	142	1.0×10^9
Second profile		16×10^9
Third profile		12×10^9

SNOW SURVEY

Pits were dug in the snow on the main Taku and its branches to determine the depth of the 1952–1953 snow cover and to determine variation of density with depth. Pits were dug about 16 miles from the terminus at an elevation of 3550 feet and about 28 miles up the main branch of the Taku at an elevation of about 4700 feet (Fig. 1). A 15-foot prod rod made from quarter-inch copper tubing was used at the pit sites to try to correlate the ice bands in the pits with the resistance encountered by the prod as it was forced down through the snow. On August 2 at the first pit the prod

encountered very hard ice bands at 6 feet 6 inches to 7 feet 6 inches and again at about 11 feet 2 inches. In the pit the bottom of the 1953 snow cover (snow that fell during the 1952–1953 winter) was found between 6 feet 8 inches and 7 feet. At this depth an ice band about 2 inches thick—thicker than the bands above—was found. It also differed from the other ice bands in that it was made up of about a half a dozen bands cemented together to give it a laminated appearance. The distinctly dirty appearance of parts of this ice band indicated that it marked the bottom of the 1953 snow cover. At 11 feet was a very dirty ice band with continuous ice below it; this marked the bottom of the 1952 snow cover. Thus, at this pit the 1953 snow depth on August 2 was 6 feet 9 inches, and the depth remaining from the 1951–1952 season was 4 feet 3 inches. In this area the snow cover was completely removed during the summer of 1951. In 1952 Edward LaChapelle estimated from pit studies that 4 feet 4 inches of snow having a density of 0.52 was left at the end of the ablation season. His estimate agrees well with the 1953 measurements.

At the second pit site the prod rod hit ice bands that gave considerable resistance at depths ranging from 10 feet 6 inches to 11 feet and again at about 13 feet 6 inches and also at about 17 feet 6 inches. It was assumed that the depth of the 1953 snow cover was approximately 10 feet 6 inches as the prod encountered considerably more resistance below this level than above it. At about 10 feet 8 inches, there was a fairly large ice band but no dirty layer which would have given the required proof that this depth marked the bottom of the 1953 snow cover. The results of other prod readings made over widely scattered areas of the Taku Glacier suggest, however, that the 1953 snow depth on July 31 was 10 feet 8 inches at this pit.

The snow studies were made at several dozen locations on the main Taku and its branches (Fig. 1). The depth of the snow was determined by use of a 15-foot prod rod. In ideal cases the bottom of the 1953 snow cover was easily detected when the prod hit a hard ice band and the snow below this ice band offered much more resistance to further prodding. In places, because more than one prominent ice band was encountered, the true snow depth was uncertain. By working out from the first pit site and other points where true snow depth was known, however, it was usually possible to determine which ice band marked the 1953 level. A more serious problem was encountered

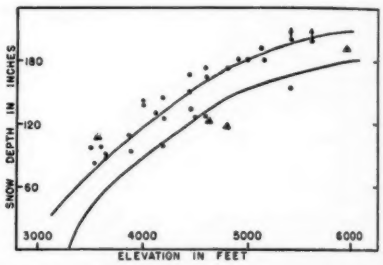

FIGURE 5.—DEPTH OF SNOW AS A FUNCTION OF ELEVATION ON THE TAKU GLACIER

Upper curve is for August 1, 1953; lower curve is estimated depth at end of 1953 ablation season. Triangular points are depths as found in pits about August 1, 1952.

at high elevations where the snow depth was greatest and the ablation season shortest. In these important regions the ice bands and demarcation characteristics became less pronounced. In addition the depth of the snow became so great that the bottom could not be reached by the prod. Thus, little is known about the highest accumulation areas except that the snow cover was more than 17 feet or the line between the 1952 and the 1953 seasons was too indistinct to be detected.

The density of the snow was also measured at each location about a foot below the surface. The density measured throughout the depths of the pits and over the whole area was surprisingly constant—between 0.50 and 0.60, averaging about 0.55 grams per cubic centimeter.

Inasmuch as the snow survey was made about August 1, which is before the end of the ablation season, extrapolations must be made to determine the depth of snow at the end of the ablation season because this is the important value. This computation is fairly simple if our values and the values of previous parties are used for rates of ablation and the length of the ablation season. Figure 5 shows the snow depth as a function of elevation. The upper curve is an average of the results found on August 1.

The lower curve gives the estimated depth at the end of the 1953 ablation season for snow of density 0.55. The triangular marks give the depths found by Edward LaChapelle on August 1 from pit studies made in 1952 (LaChapelle, 1954, p. 28). The elevation of the firn line at the end of the ablation season gives the point on the lower extrapolated curve where the net snow depth is zero. At an elevation of 3600 feet the ablation season ends about September 10. The rate of ablation over this period is about 1.3 inches per day. At the highest elevations the ablation season ends in late August. In view of the possible errors in snow-depth determinations, the extrapolated curve has been drawn with more emphasis on the 1952 pit depths, which are certainly quite accurate. The depths in 1952 and in 1953 are almost equal. A few crevasse studies (Miller, 1953) made a few miles above the firn line indicate that these values are a little above the average net accumulation over about the past 12 years. Both the 1952 and 1953 snow covers are probably somewhat larger than normal, however, and in the values for the accumulation the errors may be at least 30 per cent.

TABLE 4.—SNOW COVER OF THE TAKU GLACIER, 1953

Region	Snow cu. ft.	Ice cu. ft.
Main branch	27.1×10^9	16.5×10^9
North ("Matthes") branch	19.3×10^9	11.8×10^9
Northeast ("Demorest") branch	11.2×10^9	6.9×10^9
West branch	3.24×10^9	2.0×10^9
Total for Taku	60.8×10^9	37.2×10^9
Quantity above second profile		35.3×10^9
Quantity above third profile		27.5×10^9

Table 4 gives the calculated quantity of snow of density 0.55 remaining at the end of the 1953 summer as determined by assuming an average snow depth between contour intervals, measuring areas between contours, and summing. The second column gives the volume accumulated in terms of ice of density 0.90. This ice eventually moves down the Taku. All the branches contribute ice to the main stream above the first and second profiles. Only the main branch and the north ("Matthes") branch contribute ice through the third profile.

ABLATION BELOW THE FIRN LINE

Below the firn line the significant ablation is that which occurs after the previous winter's snow has been removed; the important factor is the amount of bare ice that melts or disappears each year.

Some ablation data were obtained from the tripods used to measure surface movement at the first profile. The lowering of the tripods as the ice melted was determined by triangulation. A few other readings were taken by a Juneau Ice Field party in 1949 near the terminus, and still other measurements were made about 11 miles up the glacier (Miller, 1952, p. 9, 14). Most of the 1953 information was obtained near the terminus of the Mendenhall Glacier about 18 miles away, and these data have been applied to the Taku. The two glaciers should be comparable, but some correction must be made for the slightly shorter ablation season on the lower Taku. Anthony Thomas informs the author (personal communication) that, on the basis of personal experience and the records of the Juneau Weather Bureau, the ablation season begins about the first part of May at the terminus of the Taku and ends about the end of November. The Mendenhall measurements were obtained by holes in the ice made by pounding in iron rods three-eighths of an inch in diameter to depths of 5 feet. These rods were then pulled out, and wood dowels a quarter of an inch in diameter and 4 feet in length were dropped into the holes and used as ablation stakes. During July and August the average rate of melting was 4.25 inches per day at two stations near the terminus. The rate during the first 10 days of September was 2.80 inches per day. A rate of ablation of 4.25 inches per day was assumed for June, and the rate was assumed to taper off to zero at the beginning and end of the ablation season. These data and the length of the ablation season suggest that the yearly ablation is about 50 feet per year near the terminus of the Mendenhall and about 40

feet per year near the terminus of the Taku Glacier. Figure 6 gives the estimated yearly ablation as a function of the elevation on the Taku. This graph was constructed from the rate of ablation and length of ablation season

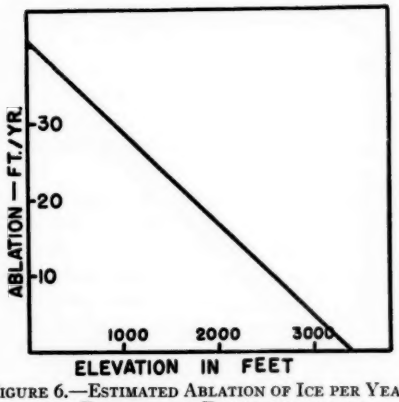

FIGURE 6.—ESTIMATED ABLATION OF ICE PER YEAR AS A FUNCTION OF ELEVATION ON THE LOWER TAKU GLACIER

TABLE 5.—ABLATION OF BARE ICE ON THE TAKU

Region	Volume cu. ft./yr.
Total below firn line	15.69×10^9
Total below first profile	4.05×10^9
Total below second profile	13.80×10^9
Total below third profile	14.77×10^9

at three locations—the zero ablation at the firn line, the estimated ablation at the terminus, and from a rough estimate based on a few ablation data obtained 11 miles from the terminus. Within the accuracy of the data, the ablation as a function of elevation was approximately a straight line. Since the errors may be as great as 25 per cent, the true ablation line may be curved.

Table 5 lists the estimated total yearly ablation of bare ice below the firn line and below various profiles of the Taku.

INTERPRETATION OF THE RESULTS

For an equilibrium or steady-state glacier the quantity of ice transported across a given cross section at or below the firn line should equal the average quantity of ice melted or removed each year below the cross section. This same volume should equal the net yearly accumulation above the firn line minus the ice melted between the firn line and the given cross section. Thus, the snow-cover data, the ablation results, and the movement studies are all related by such mathematical equations as Equations 1 through 6. Specifically, the ice moving through a given cross section can be calculated from three types of experimental data. If all three values of the volume agree, the glacier must be in equilibrium. Table 2 lists these three volumes for the three sections investigated. On the first profile the movement results and the ablation results are in fair agreement; the small discrepancy is probably due to one of the following: (1) the experimental values for the ablation are too small; (2) the Taku is still advancing slowly; or (3) the seismic depths are too large. On the second and third profiles the movement data give much smaller volumes than the ablation data. On all the profiles the accumulation as measured by the snow cover is much larger than volumes calculated from either movement or ablation. The volumes given in Table 2 have been calculated from the data given in Figures 2–6 and from Tables 1–5 by the use of equations such as 1 to 7.

These results can be explained in several ways. Since the essential data are subject to considerable errors and are not complete, it is impossible to state definitely which explanation is correct. If all the data are assumed to be reasonably accurate, then the Taku Glacier is not in a state of equilibrium. It then would appear that a series of ice waves may be advancing down the Taku. The lowest wave may be responsible for the advance of the Taku during the last 50 years and is just now being dissipated. A trough exists in the middle portion of the glacier around the second and third profiles, and another wave is building up in the upper regions of the glacier. Such an explanation is consistent with (1) the slightly greater movement near the terminus than expected for an equilibrium glacier, (2) the surprisingly small velocity of flow on the second and third profiles, (3) the much greater than expected net accumulation of snow in the higher regions.

An entirely different explanation could be given, however. There is evidence that possibly

the seismic values for the thickness of the ice on the second and third profiles may be too small. For instance, a branch of the glacier joins the main Taku about 3 miles above the second profile; seismic determinations at the junction indicate ice thicknesses of 1500+ feet. The average depth on the second profile is only 730 feet, however, whereas the maximum depth is 1033 feet according to the seismic data. Poulter's (1949, Table II) data indicate that the bed of the branch of the glacier is about 200 feet lower than the bed of the main part of the glacier at the confluence. This would appear to be very unusual.

Another reason for suspecting that the seismic thicknesses may be too small is supplied by the surface slopes of the ice. The shearing stress τ on the bottom of a glacier is (Nye, 1952a, p. 83);

$$\tau = \rho D g \sin \theta.$$

Quantity D is the thickness of the ice, ρ is the density of ice (about 0.90 g/cm^3), g is the acceleration of gravity (981 dynes/cm^2), and $\sin \theta$ is the sine of the surface slope angle θ. Glen (1953, p. 721–723) has made experiments on ice, and Nye (1952b, p. 106) and Perutz (1953, p. 934) have made calculations on other glaciers; their results indicate that at the bottom of a glacier one can expect a shear stress of about 1.0×10^6 dynes/cm^2. The experimental values for the shearing stress at the bottom of a glacier range from about 0.5×10^6 to 1.5×10^6 dynes per square cm.; an average value of 1×10^6 is reasonable and is consistent with the laboratory experiments. Table 6 lists the observed angles for the surface slopes on the three profiles. Using the above equation the expected thicknesses of the ice can be calculated; these values and the surface elevations are also given in Table 6. These calculated values are greater than the seismic values given in Table 1 for the second and third profiles but less than the average value for the first profile. The seismic values for the first profile may be too large as Poulter lists the surface elevation as about 1000 feet; whereas the recent U.S. Geological Survey topographic map (Juneau B1) and independent altimeter readings indicate the true elevation to be roughly 750 feet. As the base of the glacier is certainly at sea level or below in the region of the first profile, 750 feet would appear to be the minimum average value to be expected.

Assuming these calculated depths are correct, it is possible to recalculate the volume of ice being transported down the Taku. In Table 7

TABLE 6.—MOVEMENT PROFILES DATA

Profile	Surface slope	Surface elevation (ft.)	Calculated thickness of ice (ft.)
1	2° 50′	750	755
2	1° 14′	2725	1730
3	1° 0′	3600	2130

TABLE 7.—VOLUMES OF ICE THROUGH PROFILES

Profile	Recalculated volume (cu. ft./yr.)	Volume from ablation data
1	4.25×10^9	4.05×10^9
2	12.90×10^9	13.80×10^9
3	13.70×10^9	14.77×10^9

are given the recalculated values of the volumes determined from both the movement data and the ablation data. The two values for the volume of ice passing through the various profiles are now amazingly close; they also agree with the crude values given in Table 3. Thus, on the basis of our second explanation, both the ablation and movement data indicate that the lower half of the Taku Glacier is near equilibrium. The accumulation data are still out of line, however; either our measured values are much greater than the average snow depths over the years, or a great excess of ice is building up in the higher regions of the Taku Glacier.

Although the data presented can be interpreted ambiguously, they indicate that at least parts of the Taku are not in equilibrium. Much more ice is accumulating in the higher regions than is being dissipated below the firn line. If this trend continues, the Taku should eventually start to advance rapidly again. The time lag before excess accumulation can be noted at the terminus is not known. If the time lag is great enough, the glacier may retreat before advancing.

The insufficiency of the data for making accurate predictions means that it is important

to continue studies of mass balance in glaciers such as has been attempted on the Taku. A continuation of this work should help solve many problems on glacier regimen and behavior.

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GEOLOGY OF THE ELKHORN MOUNTAINS, NORTHEASTERN OREGON: BALD MOUNTAIN BATHOLITH

BY W. H. TAUBENECK

ABSTRACT

The Bald Mountain batholith in the Elkhorn Mountains of northeastern Oregon has intrusive relations to a heterogeneous assortment of Permian sediments and a Triassic metagabbro-serpentine sequence. The batholith is a composite intrusive composed of at least eight distinct rock types which range from norite to quartz monzonite. The different units were emplaced in a mafic to felsic sequence that commenced with norite, continued with tonalite and granodiorite, then leucogranodiorites, and ended with the emplacement of leucocratic quartz monzonite. The tonalite and granodiorite are gradational facies of the major intrusion and represent about 97 per cent of the 171 square miles of exposed batholithic rocks.

Field and petrographic evidence support the conclusion that each unit in the batholith crystallized from a discrete volume of magma. Direct evidence for stoping is limited, and the field facts imply emplacement by forceful intrusion. Areas of granitization are of restricted occurrence and regarded as by-products of magmatic activity.

Crystal fractionation of a noritic parental magma is regarded an inadequate explanation for the intrusive sequence. After the noritic magmas had crystallized, additional magmatic pulses produced the major intrusive mass and the three small leucocratic bodies. These later intrusives are regarded as originating from distinct magmas that were not sequential differentiates of noritic magma. Each later intrusion represents a disconnected magma that crystallized individually. After emplacement of the major intrusion and before emplacement of the leucocratic units, the major intrusion differentiated in place and produced a marginal tonalite that grades into a core of granodiorite.

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Introduction

The present paper is the first of three contributions concerning the geology of the Elkhorn Mountains of northeastern Oregon. Topics for two subsequent papers will be the unique Willow Lake norite and the metamorphic aureole of the Bald Mountain batholith.

Acknowledgments

This paper is a more detailed presentation of part of a thesis submitted in partial fulfillment of requirements for the Ph.D. degree at Columbia University.

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Methods

Thirty-seven weeks were spent in the field during the summers of 1950, 1951, 1952, 1953, and 1955. Approximately 80 per cent of the 400 square miles mapped during this investigation is included on the topographic sheet of the Sumpter quadrangle. This sheet provides a basis for accurate mapping. For the remainder of the area north of the Sumpter quadrangle, only inaccurate maps were available, and this mapping, especially in the vicinity of Anthony Butte, is subject to revision.

About 500 thin sections were used for petrographic study. All optic axial angles (accuracy $\pm 0.5°$) were obtained with a four-ring universal stage by direct rotation from one optic axis to the other of crystals showing acute bisectrix figures nearly centered. Refractive indices were determined by the immersion method using sodium light. After each determination, the immersion liquid was tested on an Abbe refractometer (liquids below 1.70, accuracy within ± 0.001) or a Fisher refractometer (liquids above 1.70, accuracy within ± 0.003).

The following procedure was used in extracting and studying zircons. A weighed rock sample (generally about 200 grams) was crushed to pass through an 80-mesh sieve and washed to remove dust particles. The dried material was treated with bromoform to obtain a heavy-mineral concentrate, and zircons were extracted by using a Frantz isodynamic separator. After mounting, the zircons were examined under high magnifications. Whenever possible, the lengths and widths of 300 crystals from each rock were recorded as each zircon touched the intersection of the cross hairs of the microscope during equally spaced traverses across the slides. Broken or incomplete crystals were excluded. The measured crystals were also examined to determine whether terminations were rounded or sharp.

Mineral proportions by volume for the various batholithic rocks were determined by the point-counter method (Chayes, 1949). Analyses were made of large thin sections (500–850 sq. mm), and each analysis represents at least 2000 points spaced to include most of the section. Mean and standard deviation were determined for analyses of each rock.

To afford information relative to the percentage fluctuations that might be attributed to chance or uncontrollable sectioning of the medium-grained plutonic rocks, five sections were cut from different parts of a large specimen of Anthony Lake granodiorite. Point-counter analyses of the five sections are as follows:

Potassium feldspar	8.8	7.5	6.1	7.6	7.6
Quartz	26.5	27.2	27.1	27.0	27.4
Plagioclase	53.9	53.2	54.4	57.4	55.6
Biotite	6.9	8.3	7.3	3.5	5.7
Hornblende	3.2	2.9	3.8	3.9	3.0
Accessories	0.7	0.9	1.3	0.6	0.7

Two sections were cut from each of three specimens of Elk Peak leucocratic quartz monzonite, and the analyses are as follows:

Specimen number	Potassium feldspar	Quartz	Plagioclase	Biotite	Accessories
460	19.3	32.9	44.7	2.9	0.2
460	22.1	32.6	42.4	2.8	0.1
466	28.6	36.3	33.1	1.9	0.1
466	26.0	37.3	34.9	1.5	0.3
490	26.6	36.3	35.8	1.1	0.2
490	30.8	35.6	32.7	0.7	0.2

The small variation in results is surprising and indicates that these medium-grained rocks possess a sufficient degree of internal homogeneity to be studied in reasonable accuracy by the method used.

To determine possible variations resulting from traverses that commenced at different initial points, three analyses of one section gave the following:

Potassium feldspar	8.8	9.1	9.1
Quartz	26.5	26.7	26.2
Plagioclase	53.9	53.5	53.6
Biotite	6.9	6.8	7.3
Hornblende	3.2	3.4	3.0
Accessories	0.7	0.5	0.8

Chayes (1949, p. 1) indicated that point-counter precision is better than that of the Wentworth-Hunt and Hurlbut integrators. A test by Chayes and Fairbairn (1951, p. 704–711) established the level of precision. On a

statistical basis, the theoretical analytical error and coefficient of variation (Snedecor, 1946, p. 40, 434) of the following analysis is as indicated:

	Composition	Analytical error (per cent of whole)	Coefficient of variation (index of relative error in per cent)
Potassium feldspar	7.9	0.60	7.64
Quartz	27.5	1.00	3.63
Plagioclase	52.8	1.12	2.11
Biotite	7.1	0.58	8.08
Hornblende	3.8	0.42	11.27
Opaque accessories	0.6	0.05	28.81
Nonopaque accessories	0.3	0.04	40.73

However, as the above analysis is actually the average of 32 sections of Anthony Lake granodiorite, the statistical error is as follows:

	Analytical error (per cent of whole)	Coefficient of variation (index of relative error in per cent)
Potassium feldspar	0.107	1.35
Quartz	0.177	0.64
Plagioclase	0.197	0.37
Biotite	0.102	1.43
Hornblende	0.075	1.99
Opaque accessories	0.010	5.09
Nonopaque accessories	0.007	7.20

Age Relations

In the Elkhorn Mountains the oldest rocks, the Elkhorn Ridge argillite, named by Gilluly (1937, p. 14), are a fine-grained sequence of chert, argillite, tuff, and subordinate lenses of limestone. Lava flows altered to greenstones are intercalated with these thousands of feet of sediments. The Permian (Taubeneck, 1955a) Elkhorn Ridge argillite has been intruded by an ultramafic suite that ranges from serpentinized dunite to gabbro. Investigations by T. P. Thayer (Personal communication, 1953) in the Strawberry and Aldrich Mountains of east-central Oregon indicate that the ultramafic rocks and associated gabbro are post-Permian and pre-Upper Triassic. The strongly folded Permian and Triassic rocks exhibiting low-grade regional metamorphism have been cut by the Bald Mountain batholith. Direct and indirect evidence (Taubeneck, 1955b, p. 93–96) suggests that the major intrusive unit (tonalite-granodiorite) of the Bald Mountain batholith is probably early Cretaceous.

General Character of the Batholith

Location and Topography

The Bald Mountain batholith (Fig. 1) is in the north-central part of the Elkhorn Mountains. The area south of Lat. 45° N. (Pl. 1, fig. 1) is rugged and has elevations ranging from 3500 feet to slightly over 9000 feet above sea level. Deep erosion by alpine glaciation has contributed to the formation of many excellent three-dimensional exposures well suited to batholithic studies. North of Lat. 45° N. relief is more moderate, and heavy vegetation makes this area less favorable for geologic observation.

Geographic Names

Granitic rocks of batholithic dimensions north of the mining town of Sumpter were reported by Lindgren (1901, p. 586, 655) in the first geologic investigation of northeastern Oregon. According to Hewett (1931, p. 309), Lindgren referred to these rocks as the Bald Mountain batholith and derived the name from the local designation for a prominent mountain of tonalite located about 9 miles northwest of Sumpter. In 1917 the United States Geographic Board officially designated this peak Ireland Mountain; gradually the usage of Bald Mountain was discontinued, and today the mountain is commonly called Mount Ireland. In the present paper the granitic complex will be called the Bald Mountain batholith, as this name was used by Hewett (1931).

For reference purposes certain parts of the batholith are assigned geographic names (Fig. 2). The westerly projecting mass of tonalite covering 18 square miles on the southwest is called the Monumental salient (Hewett, 1931, p. 309). The granitic area, covering 12

FIGURE 1.—LOCATION OF BALD MOUNTAIN BATHOLITH
Other batholiths shown by inverted checks.

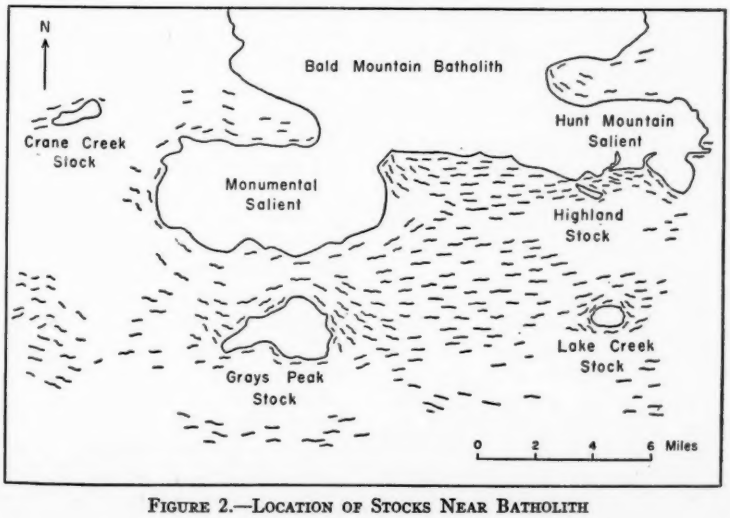

FIGURE 2.—LOCATION OF STOCKS NEAR BATHOLITH

square miles, extending eastward on the southeast corner of the batholith, is designated the Hunt Mountain salient. The wedge, predominantly metagabbro, which separates the north side of the Hunt Mountain salient from the central mass of the batholith is called the Muddy Creek wedge.

Size and Area Studied

The rocks constituting the Bald Mountain batholith crop out in an irregular area in the Sumpter and La Grande quadrangles. Approximately 171 square miles of the batholith was examined in detail, and information obtained from this area serves as the basis for most of this paper. Beyond the northern and northwestern outcrops of granitic rocks, the batholith is covered by Tertiary volcanics. Mapping (Fig. 23) was not continued north of about Lat. 45° 05′ N. because of inferior exposures and rapidly increasing cover by Tertiary volcanics. Reconnaissance covering several hundred square miles north of Lat. 45° 05′ N. revealed that restricted outcrops of granitic and prebatholithic rocks are surrounded by an almost continuous blanket of younger volcanics.

Complexity and Sequence of Intrusion

The Bald Mountain batholith is composite and contains at least eight distinct rock types which range from norite to quartz monzonite. With one exception each of the rock types represents a separate injection. Relative ages of the mapped units reveal that the general intrusive sequence was: emplacement of the norites, then tonalite and granodiorite, next leucogranodiorites, and finally quartz monzonite. This mafic to felsic sequence is commonly observed in the emplacement of composite batholiths.

Field evidence indicates that older units of batholith were entirely crystalline when the younger intrusions were emplaced. This is established by the presence of inclusions of older igneous rocks in younger intrusions, by dikes of a younger unit cutting an older unit, and by the sharp contacts between the different intrusions. This statement does not apply to the relations between the Bald Mountain tonalite and the Anthony Lake granodiorite, which are gradational facies of a single major intrusion.

The relative ages of the three mafic intrusives are unknown, and these units are arranged in the intrusive sequence according to increased silica content. This arrangement may not coincide with the relative ages of the three mafic units. According to the content of normative quartz, the Black Bear quartz gabbro is third in the intrusive sequence, but the Black Bear rock contains considerable silica introduced during metamorphism. The original Black Bear rock probably contained less silica than the Badger Butte quartz-biotite norite and possibly even less silica than the Willow Lake norite. As the original Black Bear rock was apparently the most mafic unit in the batholith, indirect evidence suggests that this unit could be the earliest of the three mafic intrusives.

Structure

Structure of Pre-Batholithic Rocks

Before the emplacement of the granitic intrusives the country rocks were tightly folded by intense orogenic forces. Deformation of orginally continuous limestone beds produced a characteristic repetition of isolated limestone blocks and lenses that, although generally offset, can be traced for distances of as much as 4 miles in the direction of regional trend. In the Elkhorn Mountains the regional trend produced by this deformation is nearly east-west, and the strata commonly dip at angles steeper than 50°. The most favorable localities for structural observations are in the glaciated mountains near the headwaters of Pine Creek and Rock Creek. In this area the bedding has been deformed into some closely appressed folds generally less than 40 feet in width. The axial planes of these nearly isoclinal folds dip steeply to the south, and the fold axes plunge to the east or west at inclinations of less than 30°. Folds of greater magnitude were not observed except along the cirque wall southwest of Rock Creek Lake where a severely deformed syncline is exposed throughout a vertical distance of 1000 feet. Although intense faulting has sheared and disrupted strata in this large

fold, the synclinal features can be recognized, especially if seen from the northeast at a distance of about half a mile. The south limb of this syncline is slightly overturned, and the axial plane dips steeply south. A more-detailed

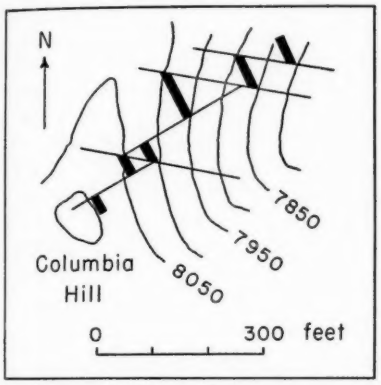

FIGURE 3.—FAULT PATTERN NEAR COLUMBIA HILL

Heavy lines (after Pardee, 1908, Fig. 5) striking N. 30° W. represent offset chert bed in Elkhorn Ridge argillite. Intrusive contact is 900 feet west of Columbia Hill.

statement of structure in the Elkhorn Mountains with regard to major axes of folding cannot be given, because deformation has been so severe and distinctive lithologies are so few that attempts to unravel the major folding of the Elkhorn Ridge argillite were unsuccessful. However, the small isoclinal-like folds observed near the headwaters of Pine and Rock creeks are undoubtedly parts of larger flexures, such as the syncline southwest of Rock Creek Lake. The steep dips that characterize most of the prebatholithic strata in northeastern Oregon suggest that much of this area has probably undergone folding that is essentially isoclinal.

One factor in the Elkhorn Mountains that makes comprehensive structural studies difficult is the widespread distributive faulting that affects the prebatholithic rocks. Faulting of this nature was first recognized by Pardee (1908, p. 88). Subsequently Pardee and Hewett (1914, p. 34) correctly interpreted these distributive faults as later than the folding but earlier than, or contemporaneous with, the emplacement of the granitic intrusives. These extensive distributive faults are structurally significant, because in many areas the fault pattern could not have originated from the forces that produced the east-west regional trend in the strata. Figure 3, a reproduction of a fault pattern shown by Pardee (1908, p. 90), is an example of a common pattern consisting of two fault systems intersecting at an angle of 40° to 60°. The acute bisectrix of this angle strikes within 20° of east-west. Such patterns can be explained only by the application of forces operating essentially at right angles to the orogenic forces responsible for the regional trend. The reason for the distributive faulting is not clear, but in the absence of any other attributable factor it seems reasonable that the compression may have been the result of forces related to the emplacement of the granitic intrusives.

The distributive faulting becomes more intense near the batholithic contacts, where nearly every joint plane shows evidence of movement. Within a mile of the contact, as on the high ridges south of the batholith, the strata are so fractured that they resemble a gigantic fault breccia. These ridge tops are 8000 feet or higher and present excellent exposures.

As the intrusive contacts are approached, the regional trend becomes increasingly disrupted. Strikes that box the compass are a common structural feature around the granitic intrusives. Such strikes can be observed best around the stocks, because large areas west and north of the batholith are covered by Tertiary volcanics that prevent structural observations in the older rocks. The disrupted regional trend around the Lake Creek stock is shown in Figure 4, and the structural trends near the Grays Peak composite stock are indicated in Figure 5. That part of the Black Mountain stock which is within the Sumpter quadrangle shows in an even more striking manner the deflection of the regional trend in the adjacent country rocks, but widespread Tertiary volcanics flanking the entire eastern margin of this stock prevent a complete structural examination.

In the northern half of the Sumpter quadrangle, to which most of this investigation is confined, schistosity in the pre-batholithic rocks is confined to the vicinity of intrusive contacts

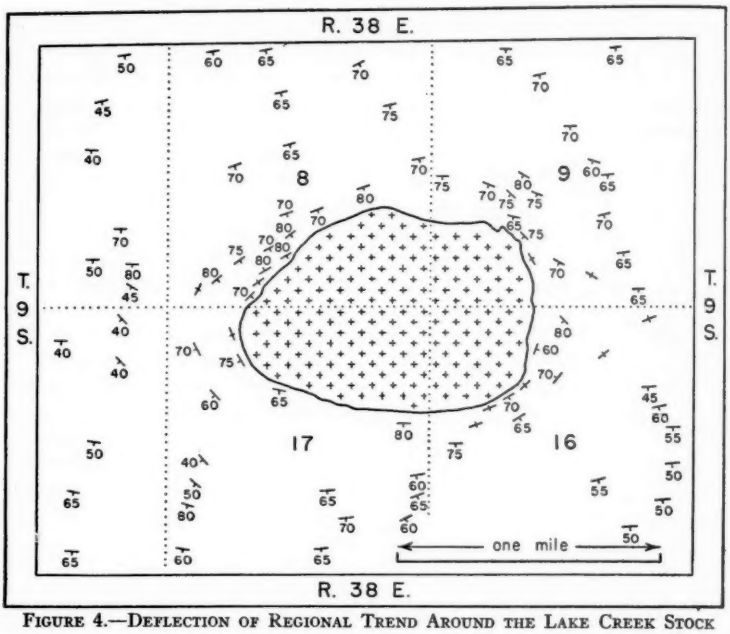

FIGURE 4.—DEFLECTION OF REGIONAL TREND AROUND THE LAKE CREEK STOCK

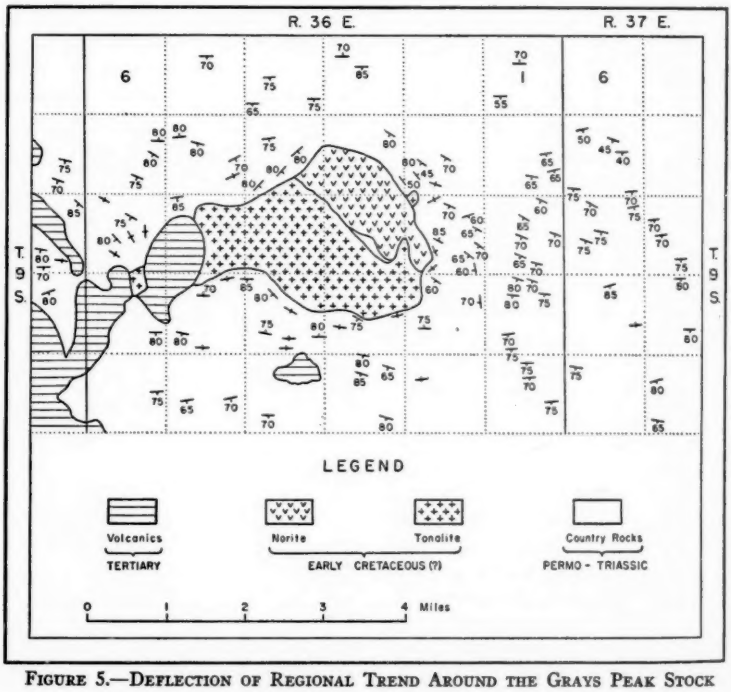

FIGURE 5.—DEFLECTION OF REGIONAL TREND AROUND THE GRAYS PEAK STOCK

and to areas of country rock that project into the batholith. The most prominent schist zone bordering the batholith is along the contact south of Red Mountain. In this vicinity the peripheral schistosity extends outward from the contact for as much as a quarter of a mile. Large wedge-shaped areas of country rock that project into the batholith become schistose as they are traced inward. For example, the wedge of Muddy Creek metagabbro that extends inward for over 3 miles is without foliation in the areas beyond the most easterly extensions of the batholith. As this wedge is followed westward into the batholith, the rock in many of the outcrops becomes a schistose amphibolite.

Structure of the Batholith

Regional structure.—The prevailing east-west regional trend of the country rock exerted some control over the emplacement of the batholith, as is shown by the shape of this composite intrusive. The Monumental and the Hunt Mountain salients are elongated in the east-west direction. Within the Hunt Mountain salient, the Willow Lake norite and the Black Bear quartz gabbro have a general east-west elongation. Even after the batholith was emplaced, the east-west direction may have been one of structural weakness, because the small leucogranodiorite intrusives within the batholith are located along an east-west trend. Not only has the regional trend played a role in governing the batholithic outline, but this trend also has been a factor in controlling the shape of nearly all the stocks (Fig. 2) associated with the batholith.

Intrusive contacts.—Wherever favorable exposures exist, the attitude of the intrusive contacts can be directly observed. Elsewhere the dip of the contact plane can be inferred if the granitic borders are characterized by either a platy parallelism or a preferred orientation of minerals or inclusions. Almost invariably the contacts dip at angles within 20° of the vertical. A vertical or slightly outward dip is most common.

West of the vicinity of the Maxwell mine, the contact dips vertically approximately 2700 feet. Within the lower 1700 feet, the contact is essentially vertical, but at higher elevations it dips steeply outward. Somewhat similar relationships are present along the entire southern boundary of the batholith, and if the arching contact were projected into space, the dome of this arch would be about 2000–4000 feet above the summits of the present granitic peaks.

In the Muddy Creek area the inward dip of the contacts is gentle. The actual contact plane cannot be seen, and its attitude is inferred from the orientation of inclusions in the tonalite. For approximately 1 mile along the north side of the Muddy Creek wedge, inclusions of metagabbro dip inward toward the tonalite at angles of 20°–45°. Elsewhere along the north side of this wedge, no evidence suggests that at depth it tapers to a point. Surface observations indicate the opposite. Summarizing this evidence of the intrusive contact planes, there is not sufficient basis for postulating that the batholith enlarges with depth or that adjacent stocks represent offshoots connected at depth with the main intrusive.

On a regional scale most of the intrusive contacts are concordant, but in detail these contacts are everywhere discordant, and within a short distance they cut across individual beds in the country rock. Where there are sharp directional changes in the outline of the intrusive, the strike of the disrupted strata is nearly at right angles to that of the contact.

Normally contacts with the country rocks are sharp and simple. Recrystallization of siliceous argillite can produce what appears from a distance to be a gradational contact, but closer examination generally reveals the point where the tonalite terminates and recrystallized country rocks commence. However, several gradational contacts were observed; probably the best one is at an elevation of 7200 feet on the southeast corner of the Lake Creek stock. Where contact breccias are present, the contact is a zone rather than a narrow line of millimeter width. Some contacts with schists are more complex, because dikes which may parallel the schistosity become numerous enough in several places to produce local alternations of igneous and metamorphic rocks.

Within the batholith all the contacts between the various members of the intrusive sequence are sharp, and although there are gradational contacts between the Bald Moun-

tain tonalite and the Anthony Lake granodiorite, these two rocks are facies of one intrusion. Except for the Black Bear quartz gabbro, igneous breccias have been observed in association with all the intrusions. Each unit of the intrusive sequence has brecciated a preceding member either along the contacts or along dikes that cut the earlier unit. Although several small extensions of the batholith are finer-grained than the adjoining intrusive mass, examples of chilled borders were not found except along the south side of the Hunt Mountain salient and in a few places around the small leucocratic intrusives.

Features of granite tectonics.—Only the more obvious structural features were noted, and no attempt was made to study the joint patterns in detail. The batholithic structures investigated were those which seemed to be most important from the standpoint of revealing possible mechanisms of emplacement.

Marginal joints and thrusts, some extending into the country rock, exist along certain parts of the intrusive contacts. These joints and upthrusts dip inward, generally at angles of 20°–40°. One of the best localities for observing marginal joints is at an elevation of 8400 feet on the ridge top between the Highland and Baisley-Elkhorn mines. Good examples of the less common marginal upthrusts occur at an elevation of 8250 feet about 1 mile south of Red Mountain. Some of the aplite-filled upthrusts have displacements of several inches, as determined from localities where these upthrusts cut across and offset inclusions in the tonalite.

Spindle-shaped to pancake-shaped clots, derived from metagabbro inclusions, are the most conspicuous linear flow structure within the batholith. Near intrusive contacts these mafic clots have been drawn out into thin and elongated streaks whose two greatest axes are parallel to the contact. A favorable locality for the study of the clots is along an irrigation ditch about a mile southeast of Bulger Hill. In this area the tonlite readily disintegrates into gruss, which has enabled the irrigation waters to cut an almost vertically walled trench as much as 30 feet deep. The walls of the trench contain numerous disk-shaped inclusions whose aspect in the third dimension can be determined by digging them out of the distingegrated tonalite. The ditch is nearly at right angles to the contact, and the progressive change in the shape of the inclusions can be observed as the channel is followed inward from the intrusive contact. For about 400 feet from the contact the inclusions are vertically drawn out in a striking manner. They are flattened triaxial ellipsoids with an elongated vertical axis. At greater distances from the contact the inclusions gradually lose their flattened form and do not possess a consistent orientation. At distances of 1000 or more feet from the contact, all inclusions are rounded and irregular.

Along most of the batholithic contacts, certain minerals have a preferred orientation. The perfection of this feature varies so that a preferred orientation is not always apparent, and in some localities it is lacking. Of the minerals in the granitic rocks, the most obvious linear parallelism is indicated by columnar crystals of hornblende. Flow lines of hornblende needles can be recognized in some rocks several hundred yards from the contact. At greater distances from the intrusive margin the linear parallelism is very faint or absent. Biotite flakes can show a preferred orientation, and less commonly tabular plagioclase crystals are aligned parallel to the contact. Conspicuous parallelism of almost vertically aligned plagioclase is present along the steeply dipping contact southeast of Hunt Mountain.

In marginal areas platy flow structures are not so common as linear parallelism. However, flow layers do exist and may be seen even in rocks from the interior of the batholith. Unlike the nearly vertical flow layers that are marginal, the platy parallelism in the interior of the batholith dips outward 25°–50°. These observations of dip are made at distances of 1–4 miles from the contact and are not sufficient in number to warrant a positive statement regarding a possible pattern for the flow layers. The present incomplete pattern suggests an arch whose apex would be near the 8775-foot peak west of Van Patten Lake. This peak is near the center of the Anthony Lake granodiorite facies which forms the core of the batholith.

Few protoclastic effects near intrusive contacts are noticeable in the field, but in places

such effects become apparent, and a distinct foliation is present. The best locality to observe protoclastic mylonite is not in the batholith but in the near-by Grays Peak stock. In this composite stock of norite and tonalite (Fig. 5), the contact between the two intrusive rocks crosses the ridge north of Grays Peak. On the crest of this ridge near the norite contact there is a 150-foot wide protoclastic zone in the tonalite. The intensity of crushing in the tonalite increases as the norite is approached. The broken hornblende and plagioclase crystals are cemented by a matrix of undeformed quartz and biotite. Considering the tectonic history of northeastern Oregon, the broken hornblende and plagioclase crystals in the marginal tonalite can be attributed only to protoclastic crushing.

Willow Lake Norite

NAME, DISTRIBUTION, AND RELATIONS: A small norite mass crops out on the high ridge north of the Baisley-Elkhorn mine. Inasmuch as many of the best exposures are along the cirque walls east of Willow Lake, the rock will be called the Willow Lake norite.

The elongated area occupied by the Willow Lake norite is divided parallel to its greatest dimension by a serrated ridge, the crest of which ranges from 7800 feet to slightly over 8500 feet. The area of the body is about a third of a square mile, and excellent exposures are present on both sides of the northeastward-trending ridge crest.

The Willow Lake norite intrudes the Elkhorn Ridge argillite, and contacts are sharp. On the north and west, the intrusive is bounded by the Bald Mountain tonalite, but few contacts are exposed, although many outcrops of these two contrasting rocks are separated by only a few feet of overburden. However, the tonalite is younger, because many dikes of tonalite (Pl. 1, fig. 2) cut the Willow Lake norite. The dikes are especially numerous near the northeast corner of the norite body.

PETROGRAPHY: The Willow Lake norite is a medium-grained rock, somewhat variable in appearance. Foliation, conspicuous near the western contacts, is present throughout much of the intrusive. In two marginal areas, there are exposures of igneous layering characterized by concentrations of hornblende in alternating layers separated by rock richer in plagioclase. The presence of these peripheral hornblende gabbros bordering the interior norite is attributed to local influx of water from "wet" country rocks before consolidation of the noritic magma. In these marginal areas, incorporated water permitted the crystallization of hornblende instead of pyroxene. Hornblende and plagioclase are the only minerals present in quantities greater than 1 per cent. Point-counter analyses of four sections at right angles to the banding gave a bulk rock average of about 62 per cent plagioclase and 37 per cent hornblende. The banding, similar to that in the San Marcos gabbro (Miller, 1938, p. 1222–1224), has local distribution, limited differences in mineral proportions between bands, and some sweeping curves. Features of the rock suggest an origin from local flow movements during a late stage in the consolidation of a magma.

A unique feature of the norite is the occurrence of a type of banding (Pl. 1, fig. 3) which has not been described previously. This unusual banding is characterized by plagioclase and mafic minerals in alternating layers in which the direction of elongation of individual crystals is essentially at right angles to the banding. The general arrangement of the crystals is similar to the crystal orientation in some epithermal veins. Crystal orientation at right angles to layering has been described in stratiform sheets, such as in the Skaergaard intrusive where the border rocks contain elongated feldspars (Wager and Deer, 1939, p. 144–150) oriented at right angles to the margin of the intrusion. Banding in the Willow Lake norite differs in the presence of alternating layers of plagioclase and mafic minerals in which the crystals of each successive layer are essentially perpendicular to the banding. Banding in the Willow Lake norite is also different from Grout and Balk's (1934, p. 881) description of hornblende needles oriented at right angles to feldspathic schlieren in the Boulder batholith. These schlieren contain no oriented minerals and are regarded as injections into early cross joints. Detailed consideration of the remarkable banding in the Willow Lake intrusion and a more complete discussion of

the norite will be postponed until more extensive field and laboratory studies are completed.

Except in the restricted areas of igneous layering where hornblende and plagioclase are the only minerals of importance, there is enough relict orthopyroxene to warrant calling the rock a norite. The plagioclase content is fairly consistent, but amounts of actinolite, hornblende, and pyroxene vary widely depending upon the degree of alteration. Modes of 14 samples of norite are given in Table 1.

TABLE 1.—MODES OF WILLOW LAKE NORITE

(In volume per cent)

Specimen number	Plagioclase	Hornblende	Actinolite	Pyroxenes	Iron ores	Apatite	Quartz	Biotite	Talc	Chlorite
72A	69.5	11.8	14.0	1.7	1.4	0.5	0.1	0.1	...	0.9
74	70.2	16.1	5.4	5.6	1.4	0.3	...	0.3	0.6	0.1
162	71.6	11.2	12.9	2.2	1.3	0.2	...	0.2	...	0.4
163	70.7	10.8	4.0	13.7	0.7	0.1	...	...	...	...
163A	73.0	4.7	11.7	6.8	1.1	0.4	...	0.1	2.1	0.1
164	71.4	16.8	9.0	1.3	1.1	0.3	...	...	...	0.1
165	68.2	20.8	6.8	0.4	2.4	0.5	0.6	0.1	...	0.2
166	73.6	12.0	4.4	7.8	1.1	0.3	...	0.8	...	...
167	75.5	14.1	6.7	0.7	0.8	0.2	0.7	1.0	...	0.3
536	60.0	32.0	5.2	0.1	1.5	0.4	...	...	...	0.8
537B	67.8	26.5	1.8	0.7	1.4	0.3	...	...	0.8	0.7
537C	66.7	18.8	1.8	9.4	2.1	0.6	...	0.1	0.5	...
538	67.6	21.4	3.5	4.5	1.3	0.5	...	0.1	0.7	0.4
539	70.3	11.3	13.7	2.6	1.5	0.4	...	...	0.1	0.1
Mean	69.72	16.31	7.21	4.11	1.36	0.36	0.10	0.20	0.34	0.29
S. D.	3.73	7.14	4.31	4.08	0.45	0.14	0.24	0.31	0.59	0.31

Plagioclase (An_{56-49}) occurs as subhedral crystals which commonly show considerable fracturing. Unlike plagioclase of some norites, the crystals are almost free from minute dust-like particles, but larger opaque inclusions are present. The plagioclase has widespread mottling which apparently represents an internal readjustment of the original crystalline structure. The mottled crystals are characterized by rounded but irregular patches of a more sodic plagioclase; this mottling is regarded as a metamorphic phenomenon possibly attributable to the emplacement of the Bald Mountain tonalite.

Hypersthene ($2V_x$: 58°, γ: 1.707) is present as slightly rounded subhedral crystals. A faint pleochroism is characteristic. Thin exsolution lamellae (Poldervaart and Hess, 1951, p. 484) of clinopyroxene are present, but schiller structure and oriented rods were not observed. Some talc occurs as an alteration product along fractures in the hypersthene.

Augite ($2V_x$: 53°, γ: 1.711) is nonpleochroic and almost colorless. Alteration of pyroxenes makes a precise determination of the hypersthene-augite ratio difficult, but examination of the freshest rocks indicates that hypersthene is about twice as abundant as augite.

Greenish-brown hornblende ($2V_x$: 79°, γ:

PLATE 1.—TOPOGRAPHY AND OUTCROPS

FIGURE 1.—Looking west from near Gunsight Peak at granodiorite peaks in central part of Elkhorn Mountains.

FIGURE 2.—Angular fragments of Willow Lake norite in a dike of Bald Mountain tonalite.

FIGURE 3.—Unique banding in Willow Lake norite. Crystals of individual bands are oriented approximately at right angles to banding.

FIGURE 1

FIGURE 2

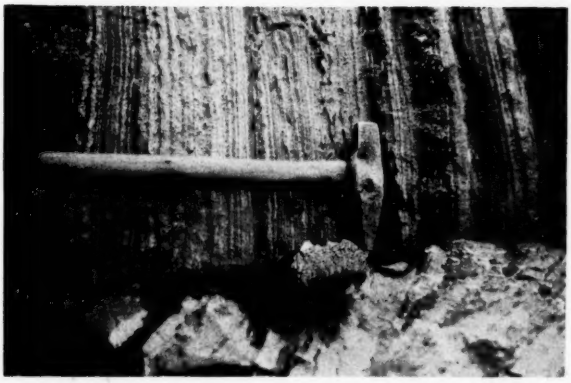

FIGURE 3

TOPOGRAPHY AND OUTCROPS

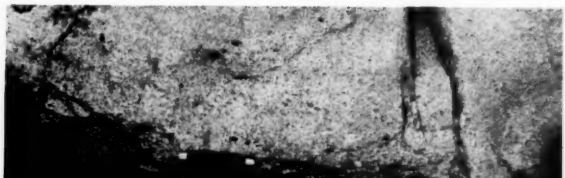

FIGURE 1

FIGURE 2

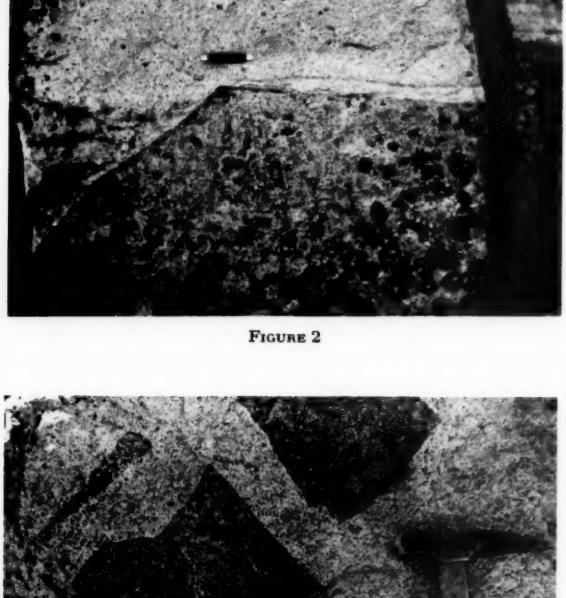

FIGURE 3

CONTACT RELATIONS

1.671) is present as conspicuous poikiloblastic crystals which almost invariably contain pyroxene relics. The amphibole grew outward from areas of pyroxene and replaced pyroxene and, to a lesser extent, plagioclase. Abundant evidence indicates a postconsolidation origin for the hornblende. Fractured plagioclase is surrounded by undeformed hornblende, and one thin section is diagonally traversed by a fracture along which a 0.1 mm wide stringer of hornblende is almost continuous. Narrow extensions of hornblende along this fracture are in optical continuity with adjacent and larger hornblende crystals through which the fracture can be projected.

Actinolite occurs as an alteration product of pyroxene. Although actinolite is present in all slices, it ranges from a minor constituent to the predominant mafic mineral. Hornblende is resistant to this alteration and remains unchanged. Therefore many of the pyroxene cores have been altered to actinolite, but the surrounding hornblende has remained unaffected. Actinolite replaces pyroxene in a crude pseudomorphic fashion and less commonly extends along fractures in the rock. Small quantities of chlorite can be associated with the actinolite, but chlorite is more characteristic of minor fractures in plagioclase. Sphene and scarce epidote accompany chlorite.

Quartz is subordinate and present as interstitial crystals which are commonly fractured and have films of potassium feldspar along the fractures. Most of the biotite is confined to margins of opaque ores. Crystals of apatite are generally characterized by a brownish color which can have a tinge of light purple. Iron ores and pyrite are the opaque minerals.

Badger Butte Quartz-Biotite Norite

NAME, DISTRIBUTION, AND RELATIONS: In the area centering around Badger Butte there are outcrops of a noritic rock which is here called the Badger Butte quartz-biotite norite. This noritic mass is of unknown areal distribution, because contacts are almost entirely concealed by extensive deposits of glacial drift. About 1 square mile of norite is exposed in the vicinity of Badger Butte, and poor exposures near the drift-covered periphery suggest that the mass has a total area of almost 2 square miles. Except for Badger Butte, the area is densely timbered and below 7000 feet in elevation. These factors contribute to some of the poorest exposures of the entire batholith.

The Badger Butte quartz-biotite norite is nearly surrounded by the Bald Mountain tonalite, but isolated outcrops of highly metamorphosed country rock west and southwest of Badger Butte indicate that on the west the norite intrusive is partly separated from the tonalite by wedge of country rocks. Along the only two located contacts of the Badger Butte intrusive, the quartz-biotite norite is intruded by Bald Mountain tonalite, and within the norite mass dikes of tonalite contain fragmented pieces of norite. Hence the Badger Butte quartz-biotite norite is older than the Bald Mountain tonalite.

PETROGRAPHY: The Badger: Butte norite is medium-grained and generally shows a crude parallelism of plagioclase. As the plagioclase is dark gray this rock is much darker than the Willow Lake norite. Texturally the rock is fairly uniform, and changes in size and distribution of poikiloblastic biotite produce the most conspicuous textural deviations. Mineralogical changes attributed to metamorphism cause considerable variation in the proportions of most minerals. Modes of 10 samples of norite are given in Table 2.

Plagioclase (An_{54-46}) occurs as subhedral crystals which show mild pressure effects and slight zoning. In many parts of the Badger Butte intrusive the crystals are distinctly aligned. The plagioclase is moderately clouded by dustlike particles, probably as a result of thermal metamorphism (MacGregor, 1931). The clouding is of uniform intensity throughout

PLATE 2.—CONTACT RELATIONS

FIGURE 1.—Contact (location of knife) of Mount Ruth leucogranodiorite and Bald Mountain tonalite. Underlying tonalite has been sheared by almost horizontal movements.

FIGURE 2.—Contact of Elk Peak leucocratic quartz monzonite and Bald Mountain tonalite. Dark spots are lichens.

FIGURE 3.—Angular fragments of Bald Mountain tonalite in Elk Peak leucocratic quartz monzonite.

entire crystals. Normally the minute particles are without directional control, although some occur in rows parallel to albite twin-composition planes. Interspersed among the "dust" particles are numerous hairlike needles, oriented in three or four directions, that intersect albite twin lamellae. These needles have maximum lengths of 0.2 mm, and although too thin to exhibit birefringence colors they can be seen by using medium-power magnifications. Antiperthite was observed in two sections with potassium feldspar in the form of small aligned squares rather than elongated stringers (Williamson, 1935, Fig. 12A). Elsewhere, small quantities of potassium feldspar are scattered randomly throughout the plagioclase or are distributed along fractures in plagioclase. Mottling of the plagioclase is less severe than in the Willow Lake norite. Internal mottling of plagioclase in the Willow Lake norite probably resulted from thermal metamorphism of higher grade, which could also have cleared the crystals of their dustlike particles. The larger opaque inclusions common in plagioclase of the Willow Lake norite may represent a coalescing of former dustlike particles.

Hypersthene ($2V_x$: 54°, γ: 1.714) forms elongated subhedral prisms with thin exsolution lamellae of clinopyroxene. Schiller structure, oriented rods parallel to the *c* axis, and pleochroism of moderate intensity are characteristic. A pale-green augite ($2V_z$: 50°, γ: 1.716) is present as crystals that range from euhedral to anhedral. The relations of the two pyroxenes are uncertain, but augite crystals molded between plagioclase laths and rarely around hypersthene suggest that augite either crystallized later or had a longer crystallization period. Hypersthene is everywhere more abundant than augite, and in relatively unaltered rocks a ratio of about 3 to 1 has been recorded.

Brownish-green to green hornblende ($2V_x$: 74°, γ: 1.681) is present as poikiloblastic crystals which are considered to be postconsolidation. Quartz, plagioclase, pyroxene, and accessory minerals are enclosed by hornblende. Almost invariably the interiors of these ramifying amphibole crystals contain pyroxene relics which served as nuclei for hornblende growth. From these nuclei extensions protrude outward along grain boundaries partly or completely enclosing adjacent minerals. Other evidence that hornblende grew in an essentially solid rock is given by crushed plagioclase surrounded by undeformed hornblende which in places has extensions along plagioclase fractures. Green is the dominant color of the Badger Butte hornblende, in contrast to brown for the hornblende of the Willow Lake norite. Both hornblendes show the same textural relations and are regarded as postconsolidation. Hence in these rocks the color of hornblende is

TABLE 2.—MODES OF BADGER BUTTE QUARTZ-BIOTITE NORITE

(In volume per cent)

Specimen number	Plagioclase	Hornblende	Pyroxenes	Actinolite	Quartz	Biotite	Opaque accessories	Nonopaque accessories
336	61.7	13.5	1.8	11.0	5.5	5.3	0.8	0.4
337	65.5	10.3	3.5	10.6	4.8	4.5	0.7	0.1
349	65.3	5.2	12.7	2.8	5.5	6.4	1.4	0.7
518A	64.7	14.8	3.1	3.9	6.4	6.3	0.7	0.1
518B	63.0	10.0	1.2	1.7	15.5	7.6	0.6	0.4
518C	63.1	10.2	4.5	3.2	7.5	10.3	1.0	0.2
520	63.5	9.2	7.3	10.4	3.8	4.2	1.2	0.4
521	61.4	12.1	1.6	7.7	7.3	8.6	0.9	0.4
524	61.0	10.9	1.4	13.4	6.4	4.6	1.6	0.7
524B	62.8	10.4	7.0	8.8	3.2	5.9	1.3	0.6
Mean	63.20	10.66	4.41	7.35	6.59	6.37	1.02	0.40
S. D.	1.58	2.58	3.66	4.13	3.43	1.96	0.34	0.22

not a criterion for origin. Shand (1942, p. 416) also concluded that color is not indicative of origin.

Biotite (β: 1.653) is present as poikiloblastic crystals, which show the same textural relations as hornblende. Therefore, biotite is also considered to be postconsolidation. Where biotite attains its largest size, the ramifying crystals extend over an area of as much as 80 sq. mm.

Actinolite is always present, but the amount varies considerably from one rock to another. Many prisms of pyroxene, generally hypersthene, have been replaced by actinolite, but the original shape can be recognized. Hornblende resisted alteration to actinolite, and there are examples of poikiloblastic hornblende surrounding cores of pyroxene which can be entirely replaced by fibrous actinolite.

Quartz was the last primary mineral to crystallize and is interstitial. Dustlike particles are minor in comparison with those in plagioclase, but larger inclusions are common, and oriented hairlike needles are present. Microscopic evidence indicates that the quartz is not entirely primary. Before crystallization of hornblende the norite was fractured, and quartz stringers extend along some of these fractures. Additional quartz veinlets date from a later fracturing which seems to be more closely related to the formation of the actinolite.

Apatite is present as colorless crystals characterized by minute needlelike inclusions parallel to the *c* axis.

Many of the zircons show sharp, irregular outlines, suggesting that some of the zircons may be skeletal crystals. Similar zircons were not observed in other rocks from the Elkhorn Mountains.

The opaque minerals are pyrite, chalcopyrite, and iron ores. These minerals were crystallized late; some fingerlike projections extend along grain boundaries between plagioclase, pyroxene, and quartz. Many of the opaque ores are rimmed and less commonly enclosed by post-consolidation hornblende and biotite.

Chlorite is not common and is absent in many sections. Although it has its source in the alteration of mafic minerals, chlorite does not preferably occur with ferromagnesian minerals but is found as irregular veinlets which transverse plagioclase. Talc is uncommon and was noted in only two slices, each of which is more fractured than normal and characterized by chlorite.

Black Bear Quartz Gabbro

NAME, DISTRIBUTION, AND RELATIONS: An unusual quartz gabbro is present on Hunt Mountain. The name of Black Bear quartz gabbro is proposed for this rock, because its characteristic features are best exposed near the Black Bear mine.

This rock covers 1⅓ square miles on the southeast slopes of Hunt Mountain.

Where the Black Bear intrusive cuts the Elkhorn Ridge argillite, the brecciated contacts are sharp. A screen of country rock on the west separates the quartz gabbro from the Bald Mountain tonalite, and a timbered ravine on the north conceals the contact of the quartz gabbro with the tonalite. The eastern margin of the Black Bear intrusive is covered by alluvium. Hence there is no direct evidence for the age of the quartz gabbro in the batholithic sequence. Originally the metamorphosed quartz gabbro may have been noritic. From the degree of metamorphism shown and the relations of the noritic rocks of northeastern Oregon, the Black Bear intrusive is assumed to be older than the Bald Mountain tonalite.

PEGROGRAPHY: The Black Bear quartz gabbro is a moderately dark-colored rock which has an average grain size between 1 and 2 mm. Most hand specimens are characterized by considerable amphibole in the form of large, scattered, dark-greenish crystals with maximum lengths of about 7 mm. Quartz, biotite, and plagioclase are also recognizable in hand specimens. Modes of 7 samples of quartz gabbro are given in Table 3.

Plagioclase (An_{82-52}) is strongly zoned with oscillatory cores bordered by outer rims that become more sodic toward the margins. Some of the euhedral to subhedral crystals show brecciation and strained twin lamellae. As most of this brecciation occurred before crystallization of quartz, shattered or mildly fractured plagioclase is associated with quartz

in which strain shadows are the maximum deformational effects. Nearly every section also shows evidence that plagioclase has been actively corroded by the residual solutions from which quartz crystallized. Embayed crystals appears to represent a postconsolidation fracturing and introduction of quartz. For example, individual fractures can be traced continuously across the width or length of thin sections. These narrow fractures are sharply defined

TABLE 3.—MODES OF BLACK BEAR QUARTZ GABBRO

(In volume per cent)

Specimen number	Plagioclase	Hornblende	Augite	Actinolite	Quartz	Biotite	Opaque accessories	Non-opaque accessories
49	47.7	19.5	0.6	4.8	19.6	7.4	0.3	0.1
51	49.1	13.4	0.7	10.2	20.3	6.0	0.2	0.1
53	45.9	12.8	2.5	20.6	13.2	4.3	0.6	0.1
391	52.3	11.8	1.6	20.0	9.8	3.5	0.8	0.2
393	49.0	4.4	0.9	12.7	22.1	10.2	0.4	0.3
394	42.4	9.1	2.3	15.3	22.2	7.6	0.7	0.4
395	50.6	6.3	1.8	17.5	18.2	4.6	0.7	0.3
Mean	48.14	11.04	1.49	14.44	17.91	6.23	0.53	0.21
S. D.	3.24	5.02	0.77	5.67	4.70	2.34	0.23	0.12

of plagioclase, commonly in optical continuity with adjacent individuals, are partly or wholly surrounded by quartz.

Brownish hornblende ($2V_x$: 79°, γ: 1.670) of primary origin occurs as reaction rims around pyroxene. Most of the rims are rather narrow, but some hornblende has almost replaced entire crystals of augite. During subsequent alteration of the pyroxene to actinolite, hornblende has persisted with little or no change. Although augite ($2V_z$: 49°, γ: 1.717) is the only pyroxene now present, the shape of many actinolite aggregates suggests that hypersthene was an original constituent.

A surprising amount of quartz is present. There are places in most thin sections where part of the quartz seems to have replaced plagioclase. Less commonly, even the mafic minerals show evidence of replacement. Some of the quartz is primary and fills interstitial spaces between euhedral or subhedral plagioclase. This relationship is best observed in sections that contain a minimum of quartz. In sections with average to maximum percentages of quartz, brecciated plagioclase and ferromagnesians are healed by unfractured quartz in optical continuity with larger areas of intersertal quartz. Although part of the brecciation could be protoclastic, some of it where they cut across either plagioclase or ferromagnesians but disappear where they encounter interstitial areas of quartz. In such areas the quartz is commonly in optical continuity with that in the nearest portion of an adjacent fracture.

Actinolite has almost replaced original pyroxene. Nearly all the hornblende persists as a reaction mantle around former pyroxenes altered to a mat of fibrous actinolite. Hence many prominent amphibole crystals contain large cores of actinolite bordered by narrow strips of primary hornblende. Where it is not rimmed by hornblende, actinolite generally frays out into the surrounding minerals or penetrates them along fractures. Wherever quartz is in contact with former pyroxene, actinolite along the contact generally assumes the form of radiating needles which extend into the quartz. The presence of scattered actinolite needles in quartz-filled fractures and the occurrence of clusters of needles radiating outward from the plagioclase walls of such fractures imply that the alteration of pyroxene to actinolite either preceded or, more likely, coincided with the main crystallization period of the introduced quartz.

Biotite (β: 1.629) generally is present as small flakes widely distributed throughout the rock. Some extends irregularly along grain

boundaries between plagioclase and quartz. Shreds of biotite are scattered throughout pseudomorphic actinolite aggregates, and as no biotite occurs in the unaltered pyroxenes, the shreds apparently formed in conjunction with the formation of actinolite. These facts, together with a positive correlation between amounts of biotite and quartz content, strongly suggest that the formation of biotite, actinolite, and part of the quartz was a closely related process which was largely responsible for transforming the original Black Bear rock into a quartz metagabbro.

Apatite has two modes of occurrence, although they are not present in every section. Generally the crystals are stout prisms, but a second type of apatite, principally associated with quartz, has the form of slender rods as much as 0.5 mm in length. Some of these rods extend across boundaries of adjacent quartz crystals or project into plagioclase. This apatite crystallized late and probably grew in part by replacement in an essentially solid rock.

Tourmaline was observed in three of the seven sections of quartz gabbro. Although the small crystals are easily overlooked, they are sufficiently numerous to be noteworthy, and their presence is an additional indication that the Black Bear intrusive has been modified by emanations associated with the emplacement of the Bald Mountain tonalite. Other accessories are zircon, pyrite, and iron ores.

Bald Mountain Tonalite

NAME, DISTRIBUTION, AND RELATION: The Bald Mountain tonalite is here named for the batholith of which it is the major rock component. This rock is characteristically exposed on Mount Ireland (Bald Mountain).

The tonalite occupies approximately 131 of the 171 square miles of batholithic rocks included in this investigation. It crops out throughout the Monumental salient, most of the Hunt Mountain salient, and in a wide zone around the granodiorite core of the batholith.

The Bald Mountain tonalite intrudes the Permian sedimentary and volcanic sequence and the Triassic ultramafic-gabbroic series. The tonalite intrudes the more mafic members of the batholith and is cut by the leucocratic granodiorites and quartz monzonite. Near the center of the batholith the tonalite grades into the Anthony Lake granodiorite.

PETROGRAPHY: The Bald Mountain tonalite is a medium-grained pale-blue-gray rock in which there is a sprinkling of dark minerals. Except for variations in the biotite, the grain size is fairly uniform and averages 2–3 mm. Plagioclase, quartz, biotite, hornblende, potassium feldspar, and sphene are recognizable in hand specimens.

Attempts were made to determine over a wide area in the batholith the mineral percentages by volume of the major intrusive rock. Certainly compositional characteristics should be critically considered in any theory that attempts to explain the origin of batholithic rocks. Therefore, hand specimens were collected from nearly every square mile of tonalite, and tabulations of volume percentages of major constituents were made by point-counter analysis of 101 thin sections. These analyses indicate that differences in composition exist between the border and the interior parts of the batholith. Modes for 41 samples of tonalite less than half a mile from the contact are given in Table 4, and modes for 60 samples of tonalite more than half a mile from the contact are given in Table 5. Range and average mineral percentage by volume for the two groups of tonalites are compared in Table 6. Modes for tonalites from the different stocks (Fig. 2) are given in Table 7.

Plagioclase (An_{48-14}) is generally subhedral, and the most distinctive feature of this mineral is its zoning. Interiors of crystals exhibit delicate oscillatory zoning and are enclosed by mantles which become progressively more sodic toward the periphery. The thin shells constituting oscillations have an approximate range in composition from An_{48} to An_{32}, and as many as 26 alternating zones have been counted in a single crystal. Almost invariably the oscillations are idiomorphic, and in the few examples where corrosion is present, it is usually confined to the outer margin of the oscillatory part of the crystal. There is no systematic differential distribution of fine inclusions within the plagioclase, but if inclusions are present, they are most commonly located at or near the margin of the oscillatory core.

The oligoclase jacket differs in volume, dis-

TABLE 4.—MODES OF BALD MOUNTAIN TONALITE LESS THAN HALF A MILE FROM CONTACT
(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Hornblende	Biotite	Accessories	
						Opaque	Non-opaque
1	2.7	24.1	60.2	5.9	7.0	0.1	0.0
23	0.6	21.9	60.7	7.3	9.1	0.2	0.2
26	0.0	17.3	56.4	13.4	12.3	0.5	0.1
30	0.0	24.9	63.1	2.0	9.1	0.7	0.2
37	0.1	19.2	61.5	11.3	7.6	0.1	0.2
48	1.4	23.3	58.9	7.6	7.6	0.8	0.4
69	0.1	27.8	57.9	4.7	9.3	0.0	0.2
70	0.3	28.2	55.1	5.1	10.4	0.3	0.6
79	1.3	20.1	56.8	10.9	10.1	0.7	0.1
81	0.1	20.8	61.0	13.7	3.4	0.8	0.2
88	1.3	22.4	58.2	7.6	10.3	0.1	0.1
99	0.8	20.6	59.2	5.0	14.1	0.0	0.3
105	0.6	20.6	61.7	5.4	10.5	1.0	0.2
108	1.6	24.0	58.4	6.7	8.8	0.4	0.1
109	0.3	25.4	60.1	7.9	5.3	0.7	0.3
110	0.0	23.1	64.2	6.9	5.7	0.0	0.1
119	0.3	23.1	61.1	9.2	5.4	0.7	0.2
176	1.4	21.0	62.8	3.0	11.4	0.3	0.1
177	0.6	25.3	58.7	7.6	7.5	0.0	0.3
179	3.7	21.6	59.9	7.6	7.0	0.1	0.1
183	0.2	24.0	60.7	5.6	9.1	0.3	0.1
185	0.4	25.4	60.6	3.9	9.1	0.2	0.4
186	0.0	21.1	61.8	7.5	9.4	0.1	0.1
192	0.8	25.5	53.6	5.0	14.8	0.1	0.2
193	0.1	21.8	58.1	7.0	12.7	0.1	0.2
212	0.4	24.3	64.7	4.6	5.7	0.1	0.2
214	0.1	22.3	57.5	7.6	11.9	0.4	0.2
215	0.3	20.1	62.6	6.3	10.2	0.4	0.1
216	0.0	24.4	61.2	5.7	8.4	0.2	0.1
256	0.0	22.3	54.2	12.1	11.2	0.1	0.1
258	0.0	25.9	59.0	5.7	9.2	0.1	0.1
359	1.5	23.2	56.3	7.6	10.7	0.5	0.2
421	0.0	21.0	59.7	9.8	8.4	0.8	0.3
452	0.0	17.1	59.6	9.9	12.7	0.6	0.1
612	0.0	22.4	56.2	9.1	11.7	0.4	0.2
713	0.1	21.2	55.0	9.5	13.2	0.8	0.2
714	1.7	23.5	55.2	6.5	12.7	0.3	0.1
719	0.3	18.9	57.3	10.1	12.8	0.4	0.2
753	0.7	19.2	56.7	12.3	10.2	0.8	0.1
754	1.6	22.0	59.6	8.3	8.1	0.3	0.1
755	2.2	21.6	56.9	8.8	10.1	0.3	0.1
Mean	0.67	22.49	59.08	7.60	9.61	0.36	0.18
S. D.	0.84	2.51	2.71	2.70	2.57	0.28	0.11

TABLE 5.—MODES OF BALD MOUNTAIN TONALITE MORE THAN HALF A MILE FROM CONTACT
(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Hornblende	Biotite	Accessories	
						Opaque	Non-opaque
57	2.8	25.0	60.4	1.6	9.6	0.5	0.1
76	2.8	22.5	58.9	6.1	8.4	1.0	0.3
84	1.5	29.1	59.2	3.5	6.0	0.5	0.2
89	3.5	22.3	58.9	6.3	8.0	0.8	0.2
90	1.7	24.7	58.6	8.8	5.5	0.6	0.1
91	4.3	32.3	51.9	4.1	6.9	0.3	0.2
92	4.6	28.2	54.8	4.0	7.6	0.6	0.2
93	4.9	31.1	53.2	3.3	7.2	0.2	0.1
93A	5.5	30.1	49.8	6.3	7.4	0.6	0.3
94	6.8	30.9	50.2	3.2	8.4	0.3	0.2
95	5.4	29.9	53.3	3.4	7.1	0.6	0.3
96	4.6	25.5	56.6	5.1	7.1	0.7	0.4
97	5.1	26.6	51.8	7.5	8.2	0.5	0.3
111	10.7	25.6	52.3	6.5	3.6	0.7	0.6
114	2.5	24.2	54.9	8.4	9.0	0.8	0.2
115	4.9	24.1	53.5	8.3	8.4	0.7	0.1
121	0.8	23.0	60.7	6.6	8.1	0.6	0.2
141	4.5	31.6	54.0	2.9	6.1	0.4	0.5
144	4.1	32.1	52.6	4.4	6.2	0.4	0.2
146	5.9	31.2	54.4	2.5	5.3	0.4	0.3
147	2.3	26.4	61.7	3.1	5.7	0.7	0.1
153	4.1	29.7	54.5	2.5	8.3	0.7	0.2
154	6.6	28.6	52.5	4.5	6.7	0.9	0.2
155	3.7	24.6	59.5	4.3	7.1	0.5	0.3
158	6.8	25.9	50.8	5.7	10.0	0.7	0.1
202	1.8	27.1	57.4	3.7	9.0	0.6	0.4
239	1.6	29.1	58.8	3.5	6.0	0.7	0.3
242	1.6	26.9	60.1	4.5	5.8	0.8	0.3
244	1.5	27.0	51.1	9.9	9.5	0.9	0.1
245	3.0	28.1	59.6	1.9	7.0	0.2	0.2
246	2.1	28.8	56.7	5.8	5.6	0.6	0.4
247	4.4	24.0	56.2	5.3	9.1	0.7	0.3
249	2.1	26.8	53.6	8.1	8.9	0.4	0.1
250	0.9	25.3	57.4	6.5	9.5	0.1	0.3
251	0.4	28.2	56.0	5.6	9.4	0.1	0.3
257	3.0	25.4	54.3	7.7	8.9	0.3	0.4
264	1.7	29.6	55.8	4.4	7.7	0.5	0.3
265	1.2	25.6	59.2	5.5	7.6	0.6	0.3
266	1.5	28.1	54.6	5.0	9.6	0.9	0.3
268	1.9	31.6	55.4	3.5	6.8	0.6	0.2
269	0.9	27.0	60.2	4.2	6.5	0.8	0.4
401	2.9	23.4	52.2	8.9	11.8	0.3	0.5
453	3.8	29.7	55.3	4.0	6.7	0.4	0.1
454	2.3	26.1	56.4	6.2	8.3	0.4	0.3
455	1.7	27.2	52.9	7.1	10.4	0.5	0.2
456	1.0	25.8	58.4	7.2	7.0	0.3	0.3
492	7.4	25.4	54.2	4.6	7.5	0.5	0.4
494	1.3	24.8	56.2	9.1	8.0	0.4	0.2
495	4.0	30.3	52.6	3.9	8.5	0.5	0.2

TABLE 5—*Continued*

Specimen number	Potassium feldspar	Quartz	Plagioclase	Hornblende	Biotite	Accessories	
						Opaque	Non-opaque
613	3.7	26.1	56.2	4.2	9.0	0.6	0.2
614	4.1	27.3	54.7	5.2	7.9	0.5	0.3
710	6.7	29.8	46.1	3.9	12.6	0.7	0.2
711	4.4	23.3	60.9	5.8	5.0	0.4	0.2
712	2.1	25.8	53.3	8.6	9.2	0.7	0.3
718	6.7	31.0	54.0	2.2	5.4	0.2	0.5
725	5.1	24.2	56.2	5.2	8.3	0.6	0.4
726	2.8	30.9	52.7	5.2	7.8	0.3	0.3
759	6.8	30.8	50.1	3.2	8.7	0.3	0.2
790	4.6	28.5	55.8	3.9	6.8	0.3	0.1
792	1.8	28.8	58.9	3.2	6.6	0.6	0.1
Mean	3.55	27.38	55.38	5.16	7.74	0.53	0.26
S. D.	2.08	2.68	3.29	2.00	1.64	0.21	0.12

TABLE 6.—RANGE AND AVERAGE MINERAL PERCENTAGES BY VOLUME FOR BORDER TONALITE, INTERIOR TONALITE, AND GRANODIORITE

	Border tonalite (41 specimens)		Interior tonalite (60 specimens)		Granodiorite (32 specimens)	
	Range	Average	Range	Average	Range	Average
Potassium feldspar	0.0–3.7	0.7	0.4–10.7	3.5	4.2–14.1	7.9
Quartz	17.1–28.2	22.5	22.3–32.3	27.4	23.9–32.2	27.5
Plagioclase	53.6–64.7	59.1	46.1–61.7	55.4	47.6–58.1	52.8
Biotite	3.4–14.8	9.6	5.0–12.6	7.7	4.9–9.7	7.1
Hornblende	2.0–13.7	7.6	1.6–9.9	5.2	1.6–7.0	3.8
Accessories	0.1–1.2	0.5	0.3–1.3	0.8	0.3–1.5	0.9

tribution with respect to crystal axes, and composition. Tonalites near the batholithic contacts contain plagioclase without mantles or with narrow mantles. In the interior of the batholith the jacket is broader parallel to the *c* axis and represents a maximum of about 25 per cent by volume of the plagioclase. Within the jacket, a rapid and continuous change in composition is present, so that this part of the crystal shows no clearly separated zones. The progressive change in composition ranges from sodic andesine on the interior of the jacket to a periphery that can be as sodic as An_{14}. Inasmuch as quartz commenced to crystallize before the complete development of the oligoclase mantle, the latter in some places extends outward for short distances along grain boundaries and thus partly surrounds quartz. Most of the plagioclase does not have these somewhat irregular oligoclase borders.

Another characteristic feature of the plagioclase is the occurrence of fractured cores healed by stringers or trails of oligoclase which possess optical continuity with the oligoclase jacket. This phenomenon was probably caused by internal strain resulting from unmixing (Chao and Taylor, 1940, p. 87) of the plagioclase within the oscillatory core. During this unmixing, submicroscopic cracks were probably produced, and these passageways facilitated diffusion of more sodic plagioclase crystallizing around the crystal core. The periphery of the crystals, which did not unmix and was not fractured, does not contain these sodic trails.

Potassium feldspar occupies interstices between the other minerals. In thin sections some

potassium feldspar appears as interstitial fillings of limited distribution, but in reality it commonly forms larger poikilitic crystals which are best recognized in hand specimen by simultaneous reflections of their cleavages. Although potassium feldspar is primarily a simple filling of residual spaces, some of it owes its distribution to replacement. The optic axial angle increases toward the center of the batholith, and lowest values occur near contacts and in the Monumental and Hunt Mountain salients. The highest frequency 2V value is 51.5°, giving an approximate composition of $Or_{30}Ab_{20}$ (Tuttle, 1952, p. 557). Forty-one determinations of $2V_x$ on different crystals in 18 sections gave the range shown in Table 9.

Microcline twinning is absent from most crystals, and if present the grid pattern is very faint and restricted to only part of some crystals. String-type perthitic lamellae (Anderson, 1928, p. 149) are more conspicuous in the interior of the batholith where some lamellae are as much as 0.4 mm long. In addition to straight, narrow string lamallae, there are a few occurrences of a more irregular perthitic structure that resembles an en echelon series of small gash veins (Spencer, 1945, Pl. 6, fig. 3).

Myrmekite favors boundaries between pla-

TABLE 7.—MODES OF TONALITE FROM STOCKS

(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Hornblende	Biotite	Accessories	
						Opaque	Non-opaque
Crane Creek stock							
196	0.1	22.4	52.7	11.1	13.2	0.4	0.1
198		17.4	51.6	19.1	11.3	0.3	0.3
200	0.1	21.7	46.4	20.1	11.4	0.1	0.2
408		21.8	50.9	17.8	8.8	0.5	0.2
Mean	0.05	20.82	50.40	17.02	11.18	0.32	0.20
S. D.	0.57	2.30	2.77	4.06	1.81	0.17	0.08
Grays Peak stock							
284	0.3	23.1	47.1	16.3	12.9	0.1	0.2
805		23.6	56.6	8.1	11.5	0.1	0.1
Mean	0.15	23.35	51.85	12.20	12.20	0.10	0.15
S. D.	0.22	0.36	6.72	5.80	0.99	0.00	0.10
Black Mountain stock							
275		16.5	56.7	15.1	10.2	1.2	0.3
277		12.1	53.6	25.8	8.0	0.3	0.2
Mean		14.30	55.15	20.45	9.10	0.75	0.25
S. D.		3.11	2.19	7.57	1.56	0.64	0.10
Highland stock							
171	3.3	25.4	55.5	3.8	11.8	0.1	0.1
507		23.2	53.2	6.7	16.4	0.4	0.1
Mean	1.65	24.30	54.35	5.25	14.10	0.25	0.10
S. D.	2.34	1.56	1.63	2.05	3.25	0.22	0.00

gioclase and potassium feldspar. However, the myrmekite is not confined to such localities but is also found along boundaries where other major minerals are in contact with potassium feldspar. Although some myrmekite forms typical cauliflowerlike protuberances, such petaloid growths are not the most common form. More commonly the intergrowth occurs as a fine film which in some places extends along the entire plagioclase-potassium feldspar contact and even encloses plagioclase if the latter is surrounded by potassium feldspar. Many of the borders of corroded plagioclase are riddled with fine myrmekite, and some of the myrmekitic blebs that occur within potassium feldspar are part of corroded plagioclase relics.

Myrmekite of granitic rocks generally has been attributed (Becke. 1908, p. 377–390) to the replacement of potassium feldspar by plagioclase. For the myrmekite in the Bald Mountain tonalite, this theory is unsatisfactory. Optical evidence clearly indicates that potassium feldspar crystallized after plagioclase and that the latter had been actively corroded by residual solutions from which potassium feldspar crystallized. Additional relations to consider include the presence of myrmekite along boundaries where potassium feldspar is in contact with either biotite, hornblende, or quartz. These relations cannot be reconciled with a theory attempting to explain myrmekite as a result of replacement of potassium feldspar by plagioclase. This theory also fails to explain aggregates of myrmekite globules enclosed within potassium feldspar (Spencer, 1945, p. 93). Myrmekite in the Bald Mountain tonalite was formed as a result of some process different from that commonly used to explain myrmekitic growths. The precise nature of this process is not known, but the best correlation for the occurrence of myrmekite seems to be the presence of corroded plagioclase.

Quartz commenced to crystallize during the development of the oligoclase rims and continued to form until the rock was essentially consolidated. Near the end of the crystallization period, quartz, potassium feldspar, and and the outermost plagioclase mantles were forming simultaneously. However, textural relations indicate that the potassium feldspar crystallized after most of the quartz because the latter may show crystal faces against alkali feldspar, but the reverse relationship is seldom observed. Most quartz forms as irregularly bounded grains which exhibit incipient fracturing and shadowy extinction.

The biotite (β: 1.647–1.653) has three major habits designated as normal, poikilitic, and fine-grained. All types occur in some rocks, but in most areas in the batholith one type is dominant enough to consider it the characteristic type for that particular area.

The normal variety of biotite has a mode of occurrence expected in rocks such as the Bald Mountain tonalite. This biotite averages about 2 mm in diameter and is generally present as excellent hexagonal books. The normal biotite commenced to crystallize before either quartz, outer plagioclase mantles, or potassium feldspar and hence shows crystalline boundaries against these later minerals. Almost invariably biotite occurs independently of hornblende, and reaction rims of biotite around hornblende are rare. In some places the normal variety of biotite extends outward for short distances between grain boundaries of all the other essential minerals except potassium feldspar. Such occurrences represent the first stage in the formation of poikilitic biotite.

For some unknown reason there is a striking development of poikilitic biotite on the north side of the basin drained by the uppermost headwaters of the North Powder River. In this part of the batholith, some conspicuous crystals of biotite reach diameters of 10 mm. Although the poikilitic biotite crystallized late in the consolidation history, it was probably not post-consolidation, because the poikilitic plates enclose only plagioclase, quartz, and hornblende. Some biotite terminates against potassium feldspar, but it does not surround potassium feldspar.

The fine-grained variety of biotite is most prominent in the interior parts of the batholith. In places the biotite averages only slightly more than 0.5 mm in greatest dimension and occurs in clusters of two or more individuals. Shreds and stringers of biotite flakes occur in some small fractures located in plagioclase or potassium feldspar. This biotite is a very late mineral.

Hornblende (γ: 1.664–1.676) is of the common green variety, and most of the crystals are euhedral to subhedral. Although the crystallization period of the hornblende paralleled that of plagioclase, the greatest development of hornblende apparently took place after most of the oscillatory cores had formed and before the growth of the oligoclase mantles. After formation, many of the amphibole crystals were corroded by the rest magma from which biotite, quartz, potassium feldspar, and the oligoclase mantles ultimately crystallized. This resorption is so strong in places that only irregular or detached particles of hornblende remain. Large hornblende crystals partly enclosed by or between plagioclase retain their original shape throughout the protected part of the crystals, but some of the parts surrounded by quartz or potassium feldspar have a corroded appearance. The optic axial angle changes little except in rocks within 200 yards of the contact. Determination of 2VT on one crystal from each of 65 sections gave ranges shown in Table 10.

Iron ores (magnetite and ilmenite) are the most abundant accessories and average about 0.46 per cent of the rock. The iron ores show evidence of late crystallization. Some crystals have irregular extensions, and some even protrude in a manner resembling the protruding portions of an amoeba. Some ores replace biotite and hornblende and in both of these minerals the opaque ores extend along cleavages or fractures as offshoots from larger crystals. The iron ores tend to be concentrated along grain boundaries between other minerals, and some occur in healed fractures in plagioclase or quartz. Apparently the iron ores crystallized later than apatite and zircon. Iron ores are later than some allanite because they replace allanite where the minerals are associated. Iron ores are earlier than epidote and most of the sphene. Some of the iron ores may have crystallized early, but the evidence for replacement of biotite and hornblende, and a consistent mode of occurrence along grain boundaries and fractures, partly as irregular fingering extensions, give the impression that much of the ores crystallized late.

Apatite comprises about 0.14 per cent of the rock, and near the intrusive contacts it differs from the apatite in the interior of the batholith. The apatite near the contacts is characterized by dusty gray cores, and some of the crystals are zoned. Some large crystals show a maximum of six distinct gray rings enclosing a dark core. The intervening areas between the rings are slightly turbid or nearly colorless. Some smaller crystals show only a cloudy core. The zoned apatites are not found in the interior of the batholith, but they do extend considerable distances toward the central part of the Monumental salient. The smaller Hunt Mountain salient is essentially free from the cloudy or zoned apatite, and the only observed occurrences were in several tonalites near the contact. According to occurrence and size, the apatite in the interior of the batholith can be divided into two general types. One type forms crystals generally larger than 0.05 mm in cross section and is generally associated with hornblende, biotite, or plagioclase. The second type tends to concentrate in clusters of minute crystals along grain boundaries and fractures. These crystals are generally euhedral and do not ordinarily exceed 0.05 mm in any dimension. This apatite commonly is associated with iron ores and less commonly with sphene. The swarms of minute crystals in the interior of the batholith are rare in the marginal tonalites.

Sphene makes up about 0.07 per cent of the rocks, and although some crystals reach lengths of 1 mm most of them are much smaller. Apparently all the sphene is of late crystallization; some of it may be postconsolidation. Euhedral forms are generally associated with quartz and potassium feldspar. However, most of the crystals are irregular aggregates, elongated stringers, and interlocking meshes. Sphene prefers grain boundaries of other minerals or fractures and has a tendency to occur adjacent to either hornblende, biotite, or iron ores. Some sphene penetrates and irregularly replaces hornblende along cleavages. Several tongues from a single crystal of sphene may extend along cleavages in unaltered biotite. Sphene occurs commonly as a thin coating around iron ores. Thin stringers of sphene are present along minor fractures in plagioclase, quartz, and potassium feldspar. Under these circumstances some fingerlike stringers of sphene coalesce and form an interlocking network. Most of these minor

fractures along which sphene is concentrated formed during the transitional stage when the rock was essentially solid but while residual magmatic solutions were still available to heal the fractures. Where they cross quartz and potassium feldspar, traces of these fractures appear as "dust" trails. In the alteration of biotite to chlorite, secondary sphene was produced as a constant minor associate of chlorite.

Allanite, constituting about 0.01 per cent of the rocks, occurs as fairly well-shaped crystals commonly ½–1 mm in size. The allanite is strongly pleochroic, and some shows zoning and twinning. Nearly half of it is associated with hornblende, and there is evidence that it has replaced hornblende. Allanite is later than most of the plagioclase and seems to be earlier than quartz, potassium feldspar, and the outermost part of the plagioclase crystals. However, one example was observed where it was definitely later than quartz. Allanite is earlier than sphene and epidote. Approximately one-third of the allanite is associated with sphene molded around allanite or penetrating it along fractures. Epidote either rims allanite as an irregular coating or projects outward as small crystals from the allanite.

Zircon is found as small grains that total about 0.01 per cent of the rocks. Most crystals are euhedral and not restricted to areas of biotite but enclosed by every other major mineral. Where zircon occurs along grain boundaries it is euhedral and, if associated with iron ores or or sphene, it has crystallized first. There are no petrographic relations to suggest that zircon was anything other than an early, perhaps the earliest, mineral to crystallize.

Rutile is a minor mineral present as a sagenitic interlocking network in chloritic biotite. In nine sections, rutile (?) was observed as minute needles aligned along different directions in plagioclase. Such rutile (?) is confined to the plagioclase of a limited area of the section; some appears in only a single feldspar crystal.

Alteration products are not conspicuous in the Bald Mountain tonalite except along shear zones. Some of the plagioclase has been subjected to mild sericitization. Some biotite shows chloritization, and less commonly it has been altered to a mixture of epidote and chlorite.

Epidote occurs in one of three forms. The two most common occurrences are either as a strongly pleochroic greenish-yellow variety restricted to chloritic biotite, or as a weakly pleochroic to nonpleochroic variety in slightly altered plagioclase. Normally, neither of these varieties has crystal faces. The uncommon third variety of epidote is found as subhedral crystals filling interstitial spaces between the other minerals. This type of colorless epidote is not associated with other mafic minerals and may be a primary late magmatic mineral.

Although chlorite is minor in quantity, it has two distinct modes of occurrence. Most commonly chlorite is present as an alteration product of biotite. Chlorite is also present as fan-shaped aggregates and vermicular crystals confined generally to potassium feldspar. Rosettes and vermicular chlorite of this type are independent of biotite. Optical properties (biaxial+, $n_\gamma - n_\alpha$ = about 0.007) suggest prochlorite. In the crystallization sequence, vermicular chlorite is earlier than epidote. The chlorite is regarded as a primary mineral which crystallized before or simultaneously with potassium feldspar. No evidence suggests that this chlorite or the associated potassium feldspar formed as a by-product of the biotite-chlorite transformation (Chayes, 1955, p. 75–82). As chlorite is a more hydrated mineral than biotite, it may have formed instead of biotite in areas where the pore liquids of the rest magma were higher in water content. Chlorite has been found to be stable in the presence of excess water at higher temperatures (680°C.) than previously expected (Yoder, 1952, p. 619). Perhaps the chlorite represents recrystallization of some of the mafic elements incorporated into the rest magma during the resorption of hornblende.

Prehnite in identifiable crystals is so uncommon in occurrence that little can be stated about it. However, the prehnite characteristically fills in re-entrant angles between or adjacent to plagioclase.

Muscovite is present in very minor quantities, and the crystals are always small. It is found along fractures in quartz and fills in cavities between the other minerals.

TABLE 8.—MODES OF ANTHONY LAKE GRANODIORITE
(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Hornblende	Biotite	Accessories	
						Non-opaque	Opaque
125	12.9	27.1	50.1	1.6	7.3	0.9	0.1
126	5.0	27.6	58.1	1.8	7.0	0.4	0.1
127	10.3	25.6	55.7	2.6	4.9	0.6	0.3
128	9.4	26.1	54.7	3.1	6.0	0.5	0.2
132	5.8	29.0	53.5	3.2	7.7	0.6	0.2
133	7.6	27.6	55.1	1.8	7.3	0.4	0.2
134	6.8	28.6	51.7	6.0	5.7	0.8	0.4
135	5.7	30.3	52.0	4.3	6.9	0.7	0.1
140	14.1	25.0	47.9	5.4	6.6	0.8	0.2
142	5.6	26.1	56.9	3.0	7.8	0.4	0.2
143	4.2	30.3	53.4	3.6	7.5	0.6	0.4
145	7.5	27.0	54.9	3.4	6.3	0.6	0.3
156	11.6	26.9	51.1	3.7	6.0	0.4	0.3
157	9.6	25.8	53.0	3.2	8.1	0.3	0.0
159	6.7	28.2	50.5	5.6	7.9	0.7	0.4
160	10.0	29.6	47.6	2.9	9.3	0.5	0.1
406	10.8	32.2	47.8	2.5	5.7	0.6	0.4
458	5.4	27.0	57.8	2.8	6.4	0.4	0.2
459	8.3	26.8	52.8	1.7	9.7	0.5	0.2
460	7.5	27.7	54.0	3.9	5.8	0.7	0.4
527	8.8	28.6	52.0	1.7	8.1	0.5	0.3
528	6.7	28.8	50.4	5.7	7.5	0.5	0.4
529	7.8	31.0	50.4	3.7	6.2	0.5	0.4
530	6.2	26.8	50.9	7.2	8.1	0.5	0.3
531	5.0	24.9	54.8	5.2	8.6	0.9	0.6
532	7.3	25.6	54.2	5.6	6.3	0.6	0.4
533	7.2	27.3	52.1	4.1	8.0	0.8	0.5
715	6.1	29.2	50.4	7.0	6.1	0.7	0.5
716	10.1	27.4	51.5	2.1	8.4	0.3	0.2
717	8.8	23.9	54.6	5.3	6.6	0.5	0.3
723	7.7	30.0	51.6	4.7	4.9	0.8	0.3
724	8.4	25.3	56.8	3.3	5.8	0.2	0.2
Mean	7.97	27.60	52.76	3.80	7.02	0.57	0.28
S. D.	2.34	1.94	2.77	1.57	1.20	0.18	0.14

ANTHONY LAKE GRANODIORITE

NAME, DISTRIBUTION, AND RELATIONS: The interior of the batholith is characterized by a potash-poor granodiorite which is centrally located with respect to Anthony Lake; it is called here the Anthony Lake granodiorite.

As the Anthony Lake granodiorite has gradational contacts with the Bald Mountain tonalite, the areal distribution of the granodiorite cannot be determined precisely. In general, the Anthony Lake granodiorite commences in a zone approximately 2½ miles inward from the contacts of the batholith and therefore covers an area of about 36 square miles.

Although the Anthony Lake granodiorite probably consolidated slightly later than the Bald Mountain tonalite, the two rocks are essentially contemporaneous and represent gradational facies of the same major intrusion.

PETROGRAPHY: In most respects the petrographic description given for the Bald Moun-

tain tonalite will serve adequately for the Anthony Lake granodiorite, because the only major distinction between the two rocks is a slight difference in mineral proportions. Hand

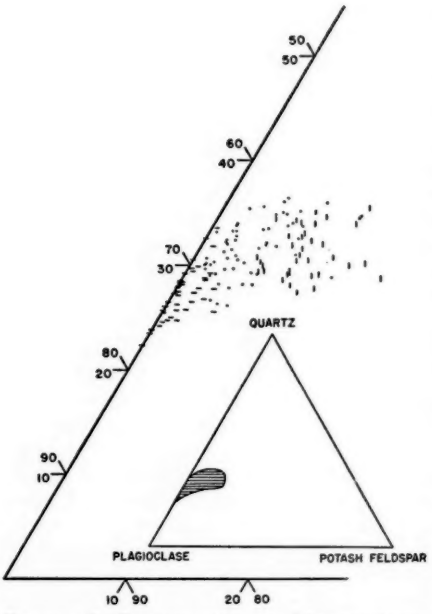

FIGURE 6.—DISTRIBUTION OF MINERALOGICAL ANALYSES OF BALD MOUNTAIN TONALITE AND ANTHONY LAKE GRANODIORITE

Shaded area of small index triangular diagram shows distribution of analyses. On enlarged portion of diagram, tonalites of the border zone are shown by horizontal dashes, the more interior tonalites by dots, and the granodiorite by vertical dashes.

specimens of Anthony Lake granodiorite show a greater content of potassium feldspar, and as a result of less hornblende, the rock has a lower color index than the Bald Mountain tonalite. Modes of 32 samples of granodiorite are given in Table 8. Mineral range and average percentage are compared in Table 6 with mineral ranges and average percentages in the border and interior Bald Mountain tonalites. For further representation, the analyses have been plotted on a quartz-potassium feldspar-plagioclase triangular diagram (Fig. 6). Specimen localities and their respective potassium feldspar content are shown in Figure 7. Figure 8 is a generalized contour map showing potassium feldspar content throughout the tonalite and granodiorite.

Under the microscope the granodiorite reveals a greater development of myrmekite than does the tonalite. It also has more prominent oligoclase mantles around plagioclase crystals. Potassium feldspar shows a continuation of the changes noted between potassium feldspar of the border and interior tonalites. Optic axial angles are higher, as is evident from Table 9 where determinations on 34 different crystals in 12 granodiorite sections are compared with 2V values of potassium feldspar in the tonalite. The range in 2V probably denotes various stages in the transformation (Tuttle and Keith, 1954, p. 68) from orthoclase to microcline microperthite. Supporting this concept is the fact that crystals which have higher optic axial angles have a shadowy extinction, and some possess fine microcline twinning. Microcline twinning was only observed in some rocks from the more central areas of the batholith where the $2V_x$ values are higher. Perthitic string lamellae are more conspicuous in the granodiorite than they are throughout the tonalite area. These facts indicate that potassium feldspar of the granodiorite has undergone a more-pronounced transformation toward the low-temperature form than has potassium feldspar of the tonalite. However, the transformation is not uniform because the optic axial angle varies considerably in several sections of granodiorite.

Plagioclase in the granodiorite has a nearly uniform compositional range, and examination of cores in 20 sections showed that the most calcic part of the cores has a constant composition of about An_{48-46}. The periphery of the oligoclase mantles is between about An_{18} and An_{12}. Excluding the tonalite within 200 yards of the contact, hornblende in the granodiorite is apparently identical to that in the tonalite, as suggested by the constancy of optic axial angles. In Table 10 determinations of $2V_x$ on one crystal from each of 31 sections of granodiorite are compared to $2V_x$ values for hornblendes of the tonalite.

MOUNT RUTH LEUCOGRANODIORITE

NAME, DISTRIBUTION, AND RELATIONS: A small mass of rather fine-grained, light-colored

leucogranodiorite is exposed on the summit and upper part of Mount Ruth and will be called here the Mount Ruth leucogranodiorite.

tonalite (Pl. 2, fig. 1) gradually varies in elevation from approximately 8300 feet to 8150 feet above sea level. On the steeper northeast

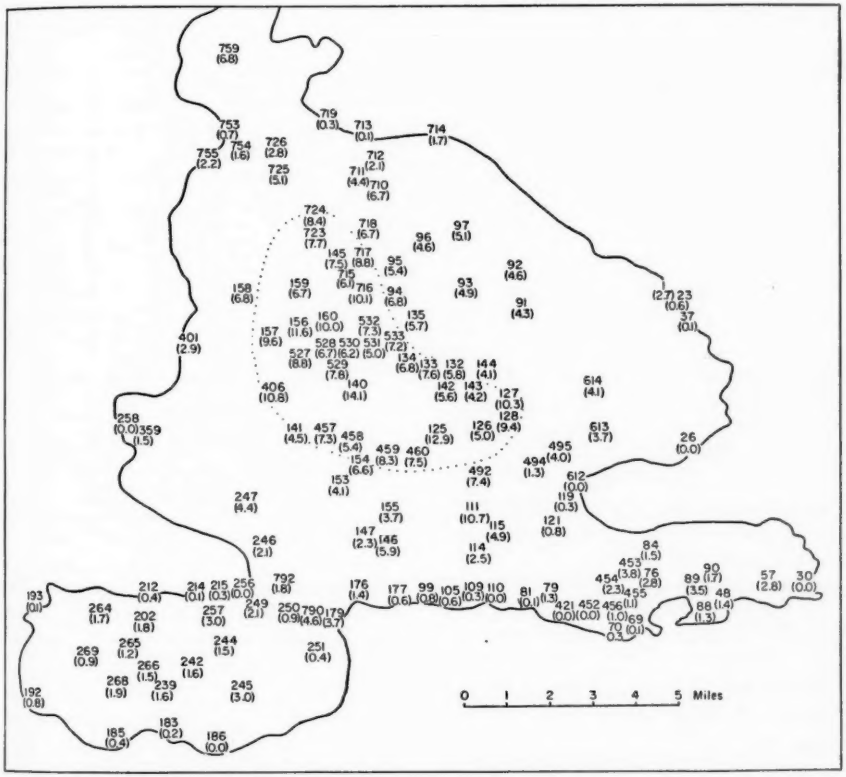

FIGURE 7.—SAMPLE LOCALITIES AND POTASSIUM FELDSPAR CONTENT OF SPECIMENS OF ANTHONY LAKE GRANODIORITE AND BALD MOUNTAIN TONALITE

Sample localities are position of decimal points of figures showing potassium feldspar contents. Area of Anthony Lake granodiorite is enclosed by dotted line.

The area occupied by this rock is about 0.06 square miles. Several dikes of similar rock type occur in near-by localities.

The Mount Ruth leucogranodiorite appears to be an almost horizontal sheetlike mass of unknown original extent, intruded into the Bald Mountain tonalite. Subsequent erosion has removed the overlying tonalite, so that the leucogranodiorite now forms the upper 250 to 400 feet of the mountain. The floor of the Mount Ruth intrusive dips to the east at about 10°. In circling the mountain the contact of the leucogranodiorite with the underlying slope, the lower contact is exposed intermittently, and for 2 or 3 feet above the contact, the leucogranodiorite may show faint foliation resulting from oriented biotite flakes which parallel the contact. At two localities the upper 6 or 8 inches of underlying Bald Mountain tonalite is strongly brecciated by almost horizontal shears, but similar relations were not observed at any other contacts. The best exposure of the contact is near the extreme northwest corner of the sheetlike intrusion, where the leucogranodiorite and tonalite are in sharp contact for about 18 feet. The nearly

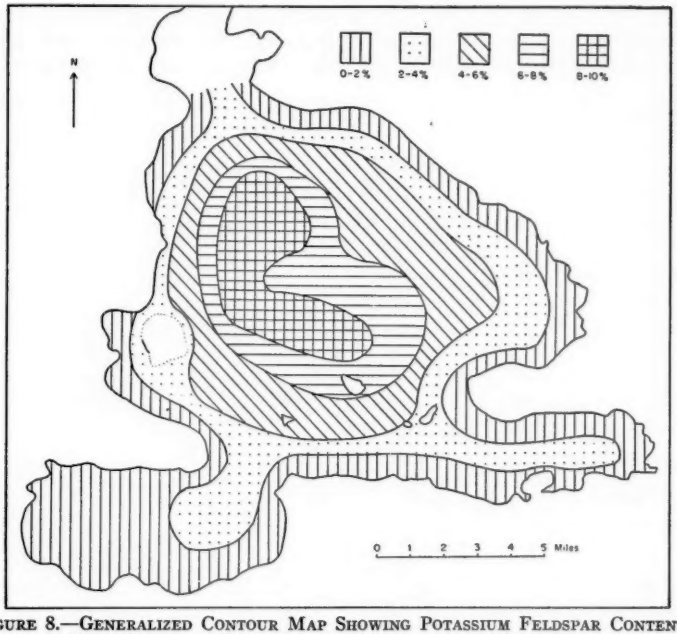

FIGURE 8.—GENERALIZED CONTOUR MAP SHOWING POTASSIUM FELDSPAR CONTENT IN BALD MOUNTAIN TONALITE AND ANTHONY LAKE GRANODIORITE

TABLE 9.—RANGE OF $2V_x$ IN POTASSIUM FELDSPAR OF TONALITE AND GRANODIORITE

Tonalite		Granodiorite	
Range in 2 V	Number of determinations	Range in 2 V	Number of determinations
40°–45°	3	40°–45°	0
45°–50°	10	45°–50°	0
50°–55°	23	50°–55°	20
55°–60°	4	55°–60°	7
60°–65°	0	60°–65°	3
65°–70°	1	65°–70°	2
70°–75°	0	70°–75°	2

whitish color of the leucogranodiorite and its textural contrast make this rock easy to distinguish from the adjacent tonalite.

PETROGRAPHY: The Mount Ruth leucogranodiorite is a somewhat porphyritic rock in which the average grain size of the groundmass ranges from 0.5 to 1.0 mm. A consistent feature of the leucogranodiorite is the presence of large scattered biotite crystals responsible for a sprinkling of dark spots in the light-gray rock. Most specimens contain several of these large crystals, which average about 4 mm diameter and generally are hexagonal. Modes for 4 samples of leucogranodiorite are given in Table 11.

Plagioclase (An_{43-14}) occurs as subhedral to anhedral crystals which vary considerably in size. Most of the laths are about 0.7 mm long, but scattered among the smaller individuals are larger crystals with maximum lengths from 3 to 4 mm. Oscillatory zoning extends outward from the andesine core to near the periphery at a composition about An_{22}. For the short distance between the last oscillatory shell and the margin, zoning is normal and becomes more sodic toward the border.

Quartz is present as euhedral to anhedral crystals generally about 0.5 mm diameter, but as is true for plagioclase and biotite larger crystals of quartz are scattered throughout the rock. Inasmuch as the crystallization period of quartz commenced before the completion of that of plagioclase, quartz can be wholly or partly enclosed in margins of plagioclase. Potassium feldspar is interstitial, having

crystallized after most of the quartz. A delicate microcline twinning is present in a few crystals but commonly is restricted to only part of the crystal. Microperthitic structures are rarely observed and, in conjunction with the micro-

TABLE 10.—RANGE OF $2V_x$ IN HORNBLENDE OF TONALITE AND GRANODIORITE

Border tonalite (less than 200 yards from contact)		Interior tonalite (more than 200 yards from contact)		Granodiorite	
Range in 2V	Number of determinations	Range in 2V	Number of determinations	Range in 2V	Number of determinations
79.5°–77.5°	3	79.5°–77.5°	0	79.5°–77.5°	0
77°–75°	5	77°–75°	0	77°–75°	0
74.5°–72.5°	5	74.5°–72.5°	0	74.5°–72.5°	0
72°–70°	4	72°–70°	14	72°–70°	2
69.5°–67.5°	0	69.5°–67.5°	25	69.5°–67.5°	22
67°–65°	0	67°–65°	9	67°–65°	7

TABLE 11.—MODES OF MOUNT RUTH LEUCOGRANODIORITE

(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Biotite	Accessories	
					Opaque	Nonopaque
148	10.4	37.1	48.3	3.8	0.3	0.1
149	7.7	35.4	52.9	3.5	0.4	0.1
151	8.5	33.5	52.1	5.4	0.4	0.1
152	9.8	36.0	50.5	3.6	0.1	0.0
Mean	9.10	35.50	50.95	4.08	0.30	0.08
S. D.	1.23	1.51	2.03	0.89	0.14	0.05

granitic texture of the groundmass, suggest that cooling in the small Mount Ruth intrusive was too rapid for appreciable exsolution. Except for scattered phenocrysts of biotite, this mineral (β: 1.648) is found as subhedral flakes and anhedral shreds. Iron ores are the most abundant accessories and are of late formation. Other accessories are apatite, zircon, rutile, sphene, and allanite.

RED MOUNTAIN LEUCOGRANODIORITE

NAME, DISTRIBUTION, AND RELATIONS: Two small bodies of leucogranodiorite are present along the top of the elongated ridge that trends northeast from the summit of Red Mountain. Both masses are petrographically alike and are here called the Red Mountain leucogranodiorite.

The two main exposures occupy an area of about 0.15 square mile and probably were connected originally. Smaller outcrops of Red Mountain leucogranodiorite are found in near-by parts of the batholith, as on Elk Peak.

The Red Mountain Leucogranodiorite intrudes the Bald Mountain tonalite, and there are strong suggestions that the two largest exposures of this rock represent disconnected parts of a sheetlike injection. The two main bodies, restricted to elevations above 8000 feet, form a capping along the great northeastern ridge of Red Mountain. A narrow saddle at about 8000 feet separates the two bodies, and this saddle probably was eroded to a level below the floor of what was once a single mass of leucocratic granodiorite.

The lower contact is generally obscured by talus, and no evidence of the original roof remains. The best-exposed contacts found are at an elevation of about 8375 feet approximately three-quarters of a mile northeast of Red Mountain summit. These contacts are distinct, because the rocks on either side differ markedly in grain size and color index. Some contacts are knife sharp, but usually they are gradational for about 1 inch. Along these merging contacts many crystals in the tonalite are fractured and healed by a fine-grained mosaic of quartz and biotite. Small biotite flakes along fracture and grain boundaries result in a darkened band which accentuates the contacts.

PETROGRAPHY: The Red Mountain leucogranodiorite is a grayish rock characterized

by scattered dark biotite clots. Except for these large biotite crystals, the grain size is rather uniform and averages 1.0–1.5 mm. Modes for seven samples of leucogranodiorite are given in Table 12.

TABLE 12.—MODES OF RED MOUNTAIN LEUCOGRANODIORITE

(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Biotite	Accessories	
					Opaque	Nonopaque
116	6.8	35.8	53.8	3.4	0.1	0.1
124	9.3	38.1	48.3	3.7	0.5	0.1
467	14.8	31.8	49.9	3.1	0.2	0.2
468	12.8	38.4	44.9	3.6	0.2	0.1
471	9.5	34.2	53.4	2.8	0.1	0.0
472	14.9	33.7	48.1	3.0	0.3	0.0
508	12.2	36.8	48.7	2.0	0.2	0.1
Mean	11.47	35.54	49.59	3.09	0.23	0.09
S. D.	3.04	2.43	3.14	0.58	0.13	0.07

Although mineral percentages in the Red Mountain leucogranodiorite are practically identical to those in the Mount Ruth leucogranodiorite, these two rocks types possess distinct differences not apparent from a comparison of mineral proportions. Texturally the Red Mountain leucogranodiorite is more uniform and has an average grain size about twice as great as that of the Mount Ruth leucogranodiorite. Alteration is more advanced in the Red Mountain leucogranodiorite, and much of the biotite (β: 1.656) is slightly chloritized. The plagioclase (An_{42-16}) in the Red Mountain leucogranodiorite is less strongly zoned and has a smaller range in composition. String-type lamellae are well developed in potassium feldspar of the Red Mountain rock but are rarely observed in the Mount Ruth leucogranodiorite. Hand specimens of the two rock types are easily distinguished, and as there is no possibility that these leucocratic rocks were once part of a single intrusive, they have been treated separately.

ELK PEAK LEUCOCRATIC QUARTZ MONZONITE

NAME, DISTRIBUTION, AND RELATIONS: Part of Elk Peak is formed by a rather small intrusive mass of leucocratic quartz monzonite, and hence this rock is called here the Elk Peak leucocratic quartz monzonite. It is present in an area covering a third of a square mile on the summit and steep southeast slopes of Elk Peak.

The leucocratic quartz monzonite has intrusive relations to the Bald Mountain tonalite-Anthony Lake granodiorite transitional zone. Sharp contacts are present (Pl. 2, fig. 2), and there are many igneous breccias (Pl. 2, fig. 3). These relations are best observed near the summit of Elk Peak, where nearly horizontal offshoots from the stocklike leucocratic intrusion extend into adjacent parts of the batholith. The smaller offshoots are noticeably finer in grain size, which suggests chilling. A small mass of Red Mountain leucogranodiorite is present in a restricted area along the northwest margin of the intrusive, and this rock has also been shattered and cut by the leucocratic quartz monzonite.

PETROGRAPHY: The slightly pinkish rock has a rather uniform grain size that averages about 1.5 mm. Plagioclase (An_{38-13}) has mild oscillatory zoning in a core surrounded by a wide mantle of progressively more sodic oligoclase. Although plagioclase does not usually enclose other minerals, it does tend to have irregular borders. After the plagioclase cores had formed, quartz, potassium feldspar, biotite, and the remaining plagioclase crystallized simultaneously, so that all these minerals normally occur as subhedral to anhedral crystals. The potassium feldspar is microcline microperthite; although some string and patch perthite is present, vein perthite is the most prevalent. Many individual crystals have delicate microcline twinning which may not prevail throughout entire crystals. In sections with a high content of potassium feldspar, some albite rims occur between potassium feldspar and plagioclase or between adjacent potassium feldspars, but myrmekite is of minor occurrence. The optic axial angle changes considerably, and shadowy extinction is common. Determinations of $2V_x$ on 56 different

potassium feldspar crystals gave the following range:

Range in 2 V	Number of determinations
50–55	19
55–60	25
60–70	7
70–80	3
80–85	2

The maximum frequency value of 2V is about 56°, which suggests a composition of $Or_{74}Ab_{26}$ according to Tuttle's (1952, p. 557) curves. Various considerations eliminate using Tuttle's curves for compositional determinations of potassium feldspars in rocks such as the Elk Peak quartz monzonite. Delicate microcline twinning is present in parts of many crystals, which indicates that the potassium feldspar is transitional to microcline. However, the only curve intersected by a 2V value of 56° is the orthoclase curve; this gives a composition too high in Ab. The composition trend from orthoclase to microcline is not along the orthoclase curve of Tuttle but in a direction almost at right angles. Potassium feldspars in batholithic rocks more felsic than the tonalite should contain smaller amounts of Ab. As Tuttle recognized (1952, p. 557), his curves cannot be used for precise determination of composition, and the possibility of gradational forms between the different series (Tuttle, 1952, p. 564) eliminates correlation of optic axial angles and composition. Although approximate compositions for certain potassium feldspars can be obtained from Tuttle's curves, the curves were apparently designed to show that alkali feldspars belong to several more or less distinct series.

Strain shadows and slight fracturing characterize much of the quartz, and small crystals of muscovite occur along some fractures. Most of the biotite (β: 1.660) is present as small and irregular flakes which may be slightly chloritized. The accessories are iron ores, sphene, apatite, zircon, and allanite. Modes for seven samples of leucocratic quartz monzonite are given in Table 13. Values for specimens 460, 466, and 490 are each average values of two thin sections.

Pegmatites

As the pegmatites have not been studied in detail, they can be described only briefly. They can be divided for descriptive purposes into relatively abundant simple pegmatites, scarce tourmaline-bearing pegmatites, and even rarer sphene-bearing pegmatites.

Table 13.—Modes of Elk Peak Leucocratic Quartz Monzonite

(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Biotite	Accessories	
					Opaque	Nonopaque
460	20.7	32.8	43.5	2.8	0.1	0.1
462	19.4	35.7	42.3	2.2	0.3	0.1
465	32.3	35.0	31.4	1.2	0.1	0.0
466	27.3	36.8	34.0	1.7	0.1	0.1
473	20.3	37.6	39.3	2.5	0.2	0.1
488	29.8	34.6	34.1	1.1	0.3	0.1
490	28.7	35.9	34.3	0.9	0.2	0.0
Mean	25.50	35.49	36.99	1.77	0.19	0.07
S. D.	5.25	1.56	4.68	0.74	0.09	0.04

Simple pegmatites are more abundant in the interior of the batholith but are scattered throughout the batholith; several occur in country rocks at short distances from the intrusive contacts. Generally the simple pegmatites are narrow dikes and not lens-shaped segregations. They are composed chiefly of quartz, feldspar, and a long-bladed dark mica. Both white and pink feldspar are present, and the field impression is that white feldspar more commonly occurs in pegmatites near the border of the batholith.

Fourteen tourmaline-bearing pegmatites were noted near the intrusive contacts; generally they are restricted to the adjacent country rocks. Although most of these pegmatites are small, an exceptionally large pegmatite is at an approximate elevation of 8000 feet about two-thirds of a mile southwest of Killamacue Lake. This imperfectly exposed pegmatite is at least 1000 feet long and perhaps 100 feet in maximum width. The most abund nt minerals

are feldspar, quartz, muscovite, and tourmaline. Some of the conspicuous crystals of tourmaline are nearly 1 foot long. Several other unidentified minerals are present in minor quantities.

TABLE 14.—MODES OF APLITE
(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Biotite	Accessories	
					Opaque	Nonopaque
13	26.6	40.8	32.3	0.2	0.1	0.0
98	33.1	38.2	28.3	0.2	0.2	0.0
150	24.8	37.6	37.4	0.1	0.1	0.0
191	31.2	35.9	32.4	0.5	0.0	0.0
252	31.2	36.1	32.3	0.4	0.0	0.0
Mean	29.38	37.72	32.54	0.28	0.08	0.0
S. D.	3.51	1.98	3.23	0.16	0.09	...

Only two of the sphene-bearing pegmatites were observed, both in country rocks at short distances from the eastern contact of the Hunt Mountain salient. The largest of the pegmatites is a dike 15 inches wide containing quartz, sodic plagioclase, hornblende, and distinctive crystals of sphene which may be nearly 1 inch in their greatest dimension.

APLITES

Aplites are far more widely distributed throughout the Bald Mountain batholith than pegmatites. Dikes of aplite are characteristically narrow, and most of them are less than 7 inches wide. Both white and pink aplites are present, but color seems to have no mineralogical correlation, as there is little variation between mineral proportions in the different aplites. Modes for five samples of aplite from widely separated areas in the batholith are given in Table 14.

LAMPROPHYRES

Only five lamprophyres were observed in the Bald Mountain batholith, and in spite of their scattered distribution all these narrow dikes have a north to northwest strike. Four lamprophyres are augite or hornblende spessartites that may show transitional tendencies toward vogesites. An olivine camptonite outcrops about 2 km north of Killamacue Lake. This camptonite contains labradorite, olivine, zoned titanaugite, analcite, and numerous alteration products. As is true for nearly all the lamprophyres in the batholith, the olivine camptonite is a highly altered rock.

OTHER DIKE ROCKS

Other dikes in the batholith can be divided into porphyries and nonporphyries.

Excluding the aplites, the most common dikes in the batholith, as well as in the country rocks, are granodiorite porphyries. These dikes reach lengths of a mile and almost invariably occupy steeply dipping shear zones with northeast strike. Phenocrysts of quartz, plagioclase, hornblende, and biotite are set in a groundmass composed chiefly of quartz and potassium feldspar. Many porphyritic dikes are strongly altered because the northeastern Oregon gold mineralization with its associated alteration is younger than the dikes in age (Hewett, 1931, p. 311) and largely restricted to shear zones along which the dikes are found. Feldspars have been sericitized, and mafic minerals have generally been altered to epidote and chlorite. Other porphyritic dikes in the batholith include quartz monzonite porphyry and rhyolite porphyry. The rhyolite porphyries were repeatedly observed along a 6-mile shear zone extending approximately N. 70° E. from the Last Chance mine.

The relatively uncommon nonporphyritic dikes are quartz monzonites, and most are confined to the south-central part of the batholith. These dikes are older than the porphyries and apparently earlier or contemporaneous with the aplites. About 800 feet north of the summit of Mount Ruth, aplites follow the same fracture occupied by an eastward-trending quartz monzonite dike, but the aplites do not cut the quartz monzonite. As the quartz monzonite dikes cut the Elk Peak intrusive and have a distinctly different type of plagioclase, the dikes are not offshoots from the Elk Peak quartz monzonite. The mineralogy of the dikes supports the conclusion that they were emplaced before the aplites and after the Elk Peak intrusive. The dikes are pink rocks of

medium grain and contain an unzoned plagioclase of about An_{20}. Modes for two samples of quartz monzonite dikes are given in Table 15.

Inclusions

Inclusions are present in all the units of the Bald Mountain batholith. The norites contain about 5 per cent of inclusions, and many of these inclusions have volumes of more than a cubic foot. Most of the inclusions in the Bald Mountain tonalite and Anthony Lake granodiorite are only a few inches in maximum dimension and represent a fraction of a per cent of the rock. Fragments of the Bald Mountain tonalite are the only inclusions in the leucocratic intrusives. The inclusions have not been studied, except those in the Bald Mountain tonalite.

Table 15.—Modes of Quartz Monzonite Dikes
(In volume per cent)

Specimen number	Potassium feldspar	Quartz	Plagioclase	Biotite	Accessories	
					Opaque	Nonopaque
491	30.6	38.3	28.9	1.9	0.2	0.1
498	30.7	40.1	28.4	0.6	0.1	0.1
Mean	30.65	39.20	28.65	1.25	0.15	0.1
S. D.	0.10	1.27	0.36	0.92	0.10	0.00

Inclusions in the Bald Mountain tonalite are plainly recognizable wall-rock xenoliths and dark inclusions whose origin is not at once evident. At least some of the dark inclusions must have been transported in a magma, because they are adjacent to cherty wall rocks of the Elkhorn Ridge argillite that could not have been the source for the inclusions. Most of the batholithic contacts are with Elkhorn Ridge argillite, and many inclusions from this formation must have been incorporated originally in the magma. However, xenoliths now visible in the tonalite include a disproportionately small number of fragments from the Elkhorn Ridge argillite. Possibly this is because most of the siliceous xenoliths (lower in the reaction series of Bowen) of Elkhorn Ridge argillite were melted or reworked more rapidly than the metagabbro inclusions, so that most of the evidence of the siliceous inclusions has been destroyed.

The dark inclusions probably have been derived from incorporated fragments of the metagabbro sequence. Amphibolite xenoliths near some amphibolite-tonalite contacts have been converted into mineralogical and textural equivalents of the dark inclusions. As the dark inclusions are distributed rather uniformly in the Bald Mountain tonalite, the magma must have moved considerably to scatter the inclusions throughout such a large mass. Most of the dark inclusions are sharply bounded, but some have gradational boundaries, and several large inclusions have a sharp contact on one side and a gradational boundary on the other. Although narrow hornblende rims about 2 mm wide were observed around several inclusions, halos are exceptional. Almost invariably the mineralogy of the tonalite remains unchanged near contacts with the inclusions. However, along several irregular batholithic contacts where xenoliths from the brecciated amphibolite wall rock show evidence of assimilation, the adjacent tonalite has a higher than normal ratio of hornblende to biotite. Possibly these occurrences indicate localities where the magma was more stagnant.

The dark inclusions contain the same minerals as the tonalite. However the dioritic inclusions contain more hornblende and biotite but much less quartz and potassium feldspar than is present in the surrounding tonalite. The inclusions may be rich in sphene and iron ores. Plagioclase, hornblende, and biotite may occur as porphyroblasts, but plagioclase porphyroblasts are the most abundant. Inclusions that show relatively little reaction with the magma have an average grain size of about 0.1 mm. Through additional reaction with the magma, the texture of the inclusions gradually becomes coarser, and thin sections reveal a local intermingling of material along contacts of the inclusions. As reaction continued, the inclusions slowly disintegrated and eventually faded almost imperceptibly into the tonalite. Near contacts, magma flowage contributed to the disintegration of inclusions, as is evident by some schlieren and poorly defined streaks

which represent inclusions in various stages of disintegration.

The evidence is clear that the tonalite has assimilated foreign rock, but the process apparently was sufficiently slow and uniform to permit the moving magma to crystallize into an essentially homogeneous rock.

TABLE 16.—FREQUENCY OF TWINNING TYPES IN PLAGIOCLASE

Rock	Type of twinning					
	Number of crystals			Percentages		
	U	A	C	U	A	C
Willow Lake norite	10	182	208	2.5	45.5	52.0
Badger Butte quartz-biotite norite	15	148	237	3.8	37.0	59.2
Black Bear quartz gabbro	27	201	172	6.8	50.2	43.0
Bald Mountain tonalite	33	204	163	8.2	51.0	40.8
Anthony Lake granodiorite	39	204	157	9.8	51.0	39.2
Mount Ruth leucogranodiorite	131	208	61	32.8	52.0	15.2
Red Mountain leucogranodiorite	104	228	68	26.0	57.0	17.0
Elk Peak leucocratic quartz monzonite	103	237	60	25.8	59.2	15.0

SPECIAL MINERALOGIC STUDIES

Plagioclase

Twinning habits of plagioclase in different intrusive units of the batholith have been studied on the universal stage. Observations were made on 400 crystals of every rock type by examining 100 crystals in equally spaced transverses on each of four sections. Twinning habits were recorded under the three categories used by Gorai (1951, p. 888–890): crystals that exhibited no twinning (U), those which showed only albite or pericline twinning (A), and plagioclase characterized by complex twinning (C). Most of the complex twins are albite–Carlsbad. The data are tabulated in Table 16.

Although no precise correlation exists between the An content and type of twinning (Fig. 9), generalizations can be made. U crystals and A twins increase in more felsic rocks, whereas C twins show a progressive decrease. Apparently this is characteristic of twinning in a series of plutonic rocks, because Gorai (1951, p. 893) has recorded a similar trend in a group of rocks in Japan.

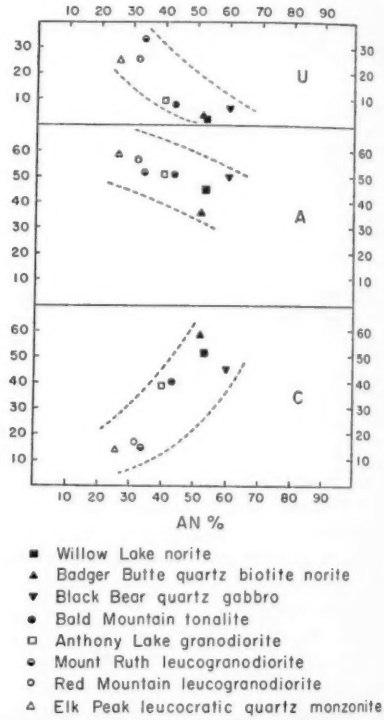

FIGURE 9.—RELATION BETWEEN AVERAGE AN CONTENT OF PLAGIOCLASE AND FREQUENCIES OF U, A, AND C TWINS

Zircon

Since the initial work of Thurach (1884, p. 203–284) an extensive literature on zircon has accumulated. The qualitative nature of papers before 1939–1942 contributed to conflicting claims. Winchell (1914, p. 128) interpreted the presence of rounded zircons in foliated rocks as evidence of sedimentary origin, whereas Armstrong (1922, p. 395) states that rounding of zircon is no criterion of origin of metamorphic

rocks. More significant claims have been made that zircon characters are uniform throughout a granitic mass (Groves, 1930, p. 235) and that zircons permit correlation of comagmatic granites (Groves, 1930, p. 232–233; Tyler *et al.*, 1940, p. 1521–1522). Stability studies of heavy minerals (Pettijohn, 1941, p. 612; Smithson, 1950, p. 14) indicate that zircon is extremely resistant in weathering and transport. Nevertheless, authigenic outgrowths of probable secondary zircon on detrital grains of this mineral have been reported by Butterfield (1936, p. 511–516), Smithson (1937, p. 281–283), and Bond (1948, p. 35–40). Although zircon in igneous rocks has generally been considered of early crystallization, Moorhouse (1953, p. 1455) considers it of late crystallization. Gillson (1925, p. 187–191) records metasomatic formation of zircon in contact metamorphism, and Wyatt (1953, p. 1496) reports that zircon of granitized rocks is recrystallized and idiomorphic. Poldervaart and Eckelmann (1955, p. 947) indicate that high proportions of zircons showing evidence of new growth may be regarded as characteristic of autochthonous granite. Smithson (1939, p. 348–360) applied statistical methods to studies of sedimentary zircons, and the method was utilized by Coetzee (1941, p. 191–193; 1942, p. 97–100) in studying zircons of igneous and metamorphic rocks. Poldervaart summarized the role of zircon in sedimentary rocks (1955, p. 433–461) and igneous rocks (1956, p. 521–554).

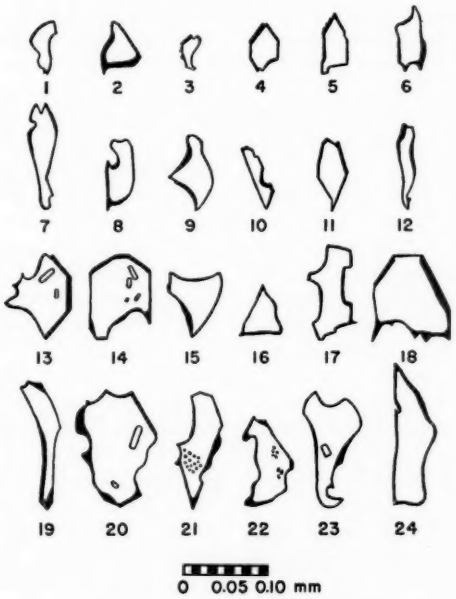

FIGURE 10.—SIZE AND SHAPE OF ZIRCONS IN WILLOW LAKE NORITE AND BADGER BUTTE QUARTZ-BIOTITE NORITE

The Willow Lake norite contains relatively few zircons in comparison with other units of the batholith, and only about 300 rather small crystals were recovered from 221 grams of crushed norite. The zircons in the Willow Lake norite are distinctive and characterized by sharp irregularites (Fig. 10). Even crystals which are somewhat rounded generally have a sharp edge somewhere along the crystal periphery. Euhedral crystals are uncommon, and many zircons do not possess any crystal faces. As

these zircons have no resemblance to sedimentary zircons, the features exhibited are regarded as skeletal growths during magmatic crystallization. Tyler *et al.* (1940, p. 1468) suggested that magmatic corrosion produced the jagged and irregular features of some zircons of Keweenawan granophyres, but these shapes probably represent early stages of growth. Jagged, sharp features are more logical in early growth than in magmatic corrosion, which should produce rounded shapes.

The Badger Butte quartz-biotite norite contains an assortment of zircons which vary markedly in shape from euhedral to extremely irregular forms. Most of the euhedral crystals are terminated by a complex assemblage of pyramidal faces that can be capped by a basal pinacoid. These euhedral crystals generally do not have length to width ratios of more than about 2:1, and crystals are somewhat similar in habit to the zircons of the Bald Mountain tonalite. Characteristically the zircons (Fig. 10) from the quartz-biotite norite form many strikingly irregular fragments, chips, slivers, or flakes. Some of the irregularities are of a shape that could not have originated by fracturing during crushing. Zircon assemblages from more silicic units of the batholith and from all the metamorphosed sediments contain few broken crystals, and as the same procedure was followed in extracting the zircons from all rocks the norites also should contain relatively few broken crystals. Therefore, these irregular pieces are regarded as zircons that exhibit incomplete growth. In support of this contention is the presence of zircons with euhedral terminations at one extremity and irregular terminations at the other. Further evidence favoring incomplete crystallization is the occurrence of every gradation between euhedral crystals and irregular fragments.

Although zircons are present in the Black Bear quartz gabbro, they are not much more prevalent than in the Willow Lake norite. Prismatic faces separated by sharp interfacial edges are common, and most crystals are relatively free from inclusions. A few of the crystals show similarities to the zircons of the Willow Lake norite, but most of the individuals are unique. The unique quality is the predominant occurrence of prisms without terminal faces at one or more generally at either extremity. Generally the extremities are irregular. Some crystals appear to be broken almost at right angles to the *c* axis, but the actual existence of such fractures is somewhat doubtful. As a test, euhedral zircons were crushed in a mortar and the fragments examined microscopically. Euhedral crystals were also fractured on a glass slide and observed during various stages of fracturing. Fractured surfaces of the test specimens did not resemble the irregular terminations of the Black Bear zircons. In addition, if the zircons of the quartz gabbro were broken during crushing, zircons of similar or larger size, especially the prismatic crystals of the leucocratic intrusives, would also have been broken. As broken prisms were not conspicuous in any other rocks from the Elkhorn Mountains, most of the zircons in the quartz gabbro probably are not broken and lack terminal faces because of incomplete growth.

Zircons were extracted from six widely separated specimens (Fig. 11) of Bald Mountain tonalite and Anthony Lake granodiorite. No perceptible difference exists in the crystals from the six rocks, as is apparent from an examination of frequency curves (Smithson, 1939, p. 349–350; Coetzee, 1942, p. 100) in Figures 12, 13, and 14. The short and stumpy crystals have first- and second-order prisms terminated by a series of pyramids. The multiplicity of pyramidal faces is characteristic, and a given order of pyramid can have as many as three different faces. The basal pinacoid is present on some crystals but generally is restricted to one extremity. Inclusions are numerous in the zircons and apparently are of random orientation.

The leucogranodiorites contain zircons that are distinctively different from those of the Bald Mountain tonalite. Frequency curves (Fig. 15) for the zircons of the leucogranodiorites have smaller width and greater elongation values. The zircons are of much simpler habit, and some crystals have a simple combination of second-order prisms terminated by first-order pyramids. The basal pinacoid is rare, and crystals contain fewer inclusions than do zircons from the tonalite. Some of the inclusions are oriented parallel to the *c* axis.

The Elk Peak intrusive contains zircons similar in morphology to those of the leucogranodiorites. However, the zircons of the Elk Peak quartz monzonite can be distinguished from those of the other leucocratic intrusives by an alteration that probably represents a stage in the transformation to metamict crystals. Many individuals are slightly altered near the ends, and some crystals with little or no birefringence are nearly opaque. In contrast to zircons from the other intrusive units, some of the Elk Peak zircons show slight zoning.

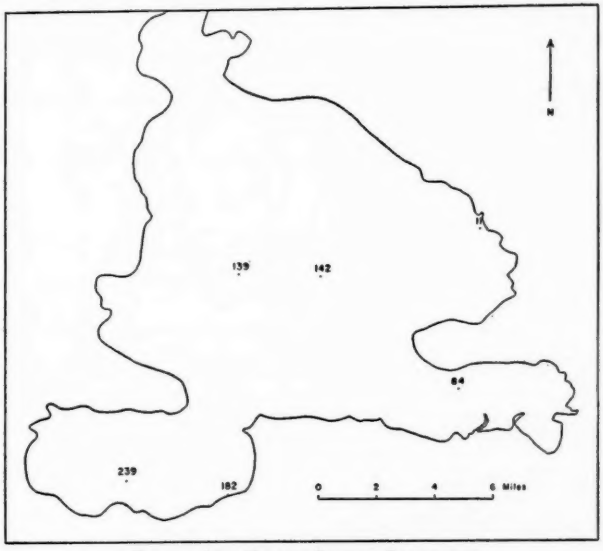

FIGURE 11.—ZIRCON SAMPLE LOCALITIES
Bald Mountain tonalite (11, 84, 182, 239) and Anthony Lake granodiorite (139, 142).

The following facts are apparent from a comparison of the zircons in the different intrusives. The quartz gabbro and norites contain relatively few zircons, most of which have crystal forms that are attributed to incomplete crystallization. Progressing from the mafic to the more silicic units of the intrusive sequence, euhedral zircons first become predominant in the tonalite. In mafic and intermediate members of the batholith, euhedral zircons are complex in habit, whereas the crystallography of zircons becomes progressively more simple in the leucogranodiorites and leucocratic quartz monzonite. As the tonalite is characterized by stubby zircons and the leucocratic units by crystals that are more prismatic, the elongation ratio increases from the intermediate to the felsic plutonic rocks.

Skeletal zircons in the norites, identical to skeletal zircons found by Duschatko and Poldervaart (1955, p. 1105) in a spilitic intrusion in New Mexico, suggest that zircon is not the earliest mineral to crystallize in basaltic magma. During the early and middle stages of fractionation of the Skaergaard magma, Wager and Mitchell (1951, p. 151) indicate that most zirconium was confined to the pyroxene lattice. In the Skaergaard differentiation sequence, zircon crystals do not appear in rocks earlier than the transgressive hedenbergite granophyres. Wager and Mitchell (1951, p. 192) note that during the later stages of fractionation, there is a marked decline in the ease of entry of zirconium into pyroxenes despite a rise in zirconium content of the liquid. These facts imply that zircons in mafic rocks are of late crystallization which would also account for their irregular and incomplete crystal shapes.

The zircon assemblage from each intrusive is characteristic of that particular intrusive.

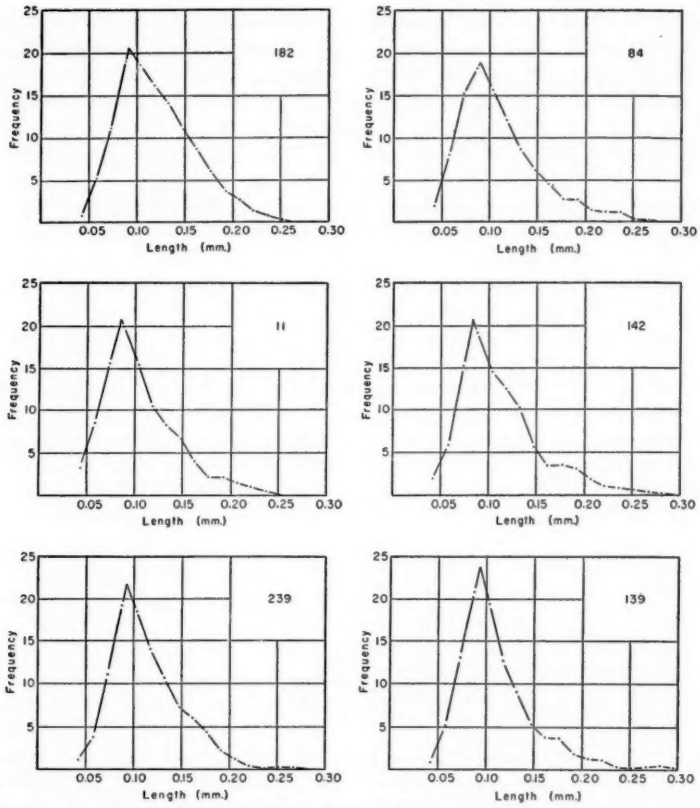

FIGURE 12.—LENGTH FREQUENCY CURVES FOR ZIRCONS OF BALD MOUNTAIN TONALITE AND ANTHONY LAKE GRANODIORITE
Specimen numbers are in upper right-hand corner.

Therefore, apparently physical and chemical conditions during crystallization of chemically dissimilar plutonic rocks are responsible for the occurrence of types of zircons characteristic of particular kinds of plutonic rocks. The distinctiveness of zircons from the different rocks is established more rigidly by statistical treatment of quantitive data of lengths and breadths using improved methods (Larsen and Poldervaart, 1955; Larsen, in preparation) better adapted to comparative evaluation of zircon data. As zircons from the norites are characteristically irregular and incomplete crystals, these zircons cannot be quantitively compared with zircons from the other intrusive units and consequently are not included. Zircon statistics computed by L. H. Larsen for samples of the more felsic intrusive rocks are given in Table 17. Figure 16 illustrates the uniformity of zircons from the major intrusive unit (tonalite-granodiorite), and the distinctiveness of these zircons from those of each of the different leucocratic intrusives. In order for two or more lines to be statistically the same, they must be essentially identical with respect to slope and position of the mean size. The consanguinity of samples 11, 84, 182, 239, 139, and 142 is indicated by the near identity of slopes and mean sizes. Sample 116 (Red Mountain leucogranodiorite) has a slope different from the tonalites-granodiorites, and the position of the mean size is quite decisive also in showing that sample 116 is of a distinctly different population. Samples 148 and 465 are

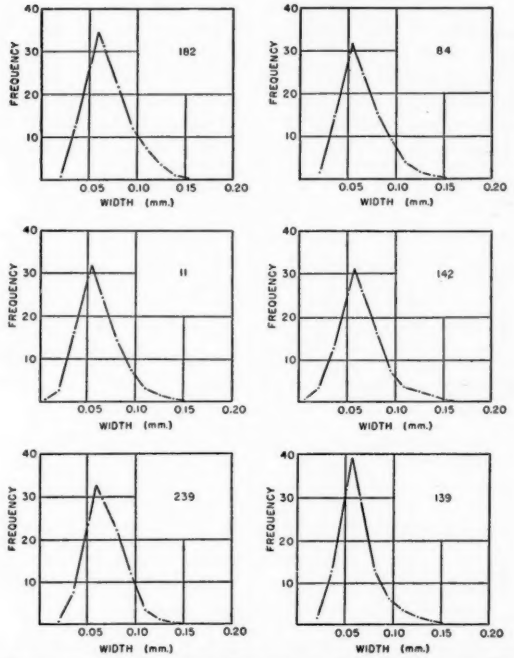

FIGURE 13.—WIDTH FREQUENCY CURVES FOR ZIRCONS OF BALD MOUNTAIN TONALITE AND ANTHONY LAKE GRANODIORITE

Specimen numbers are in upper right-hand corner.

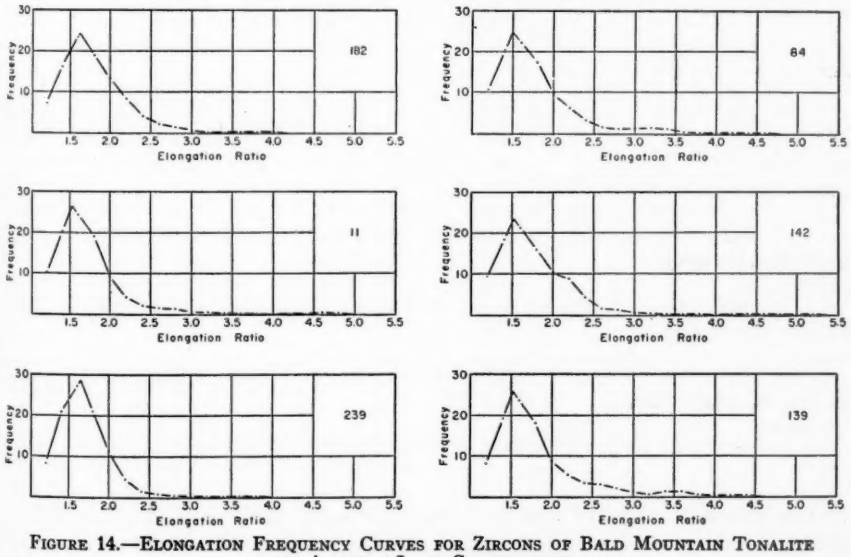

FIGURE 14.—ELONGATION FREQUENCY CURVES FOR ZIRCONS OF BALD MOUNTAIN TONALITE AND ANTHONY LAKE GRANODIORITE

Specimen numbers are in upper right-hand corner.

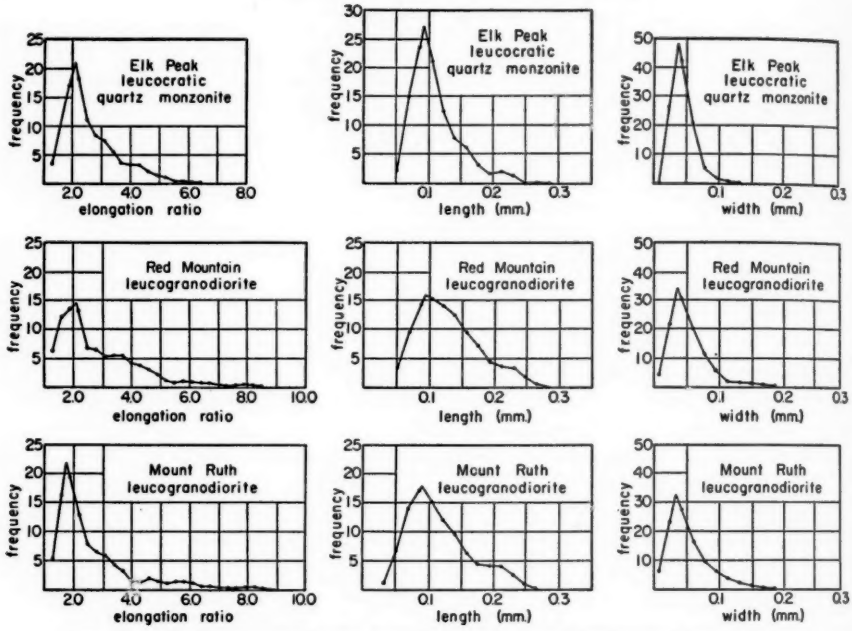

FIGURE 15.—FREQUENCY CURVES FOR ZIRCONS OF THE LEUCOCRATIC INTRUSIVES

TABLE 17.—ZIRCON STATISTICS

Sample number	Number of crystals	Tangent slope	Slope angle	Mean length (mm)	Mean width (mm)	Standard deviation, length	Standard deviation, width	Correlation coefficient
Bald Mountain tonalite								
11	300	0.5332	28° 4′	0.1059	0.0625	0.0388	0.0207	0.7187
84	300	0.4954	26° 22′	0.1109	0.0640	0.0410	0.0203	0.6984
182	300	0.4793	25° 36′	0.1242	0.0690	0.0434	0.0208	0.7233
239	300	0.5051	26° 48′	0.1140	0.0681	0.0355	0.0179	0.8091
Anthony Lake granodiorite								
139	300	0.5229	27° 36′	0.1128	0.0640	0.0406	0.0212	0.6704
142	300	0.5188	27° 25′	0.1129	0.0646	0.0443	0.0230	0.7672
Mount Ruth leucogranodiorite								
148	300	0.6401	32° 38′	0.1184	0.0535	0.0506	0.0234	0.7768
Red Mountain leucogranodiorite								
116	300	0.5075	26° 54′	0.1326	0.0528	0.0562	0.0285	0.5529
Elk Peak leucocratic quartz monzonite								
465	300	0.3567	19° 38′	0.1139	0.0451	0.0464	0.0165	0.4937

different from each other and from all other samples.

The shapes and peaks of frequency curves (Figs. 12, 13, and 14) for zircons in Bald Mountain tonalite and Anthony Lake granodiorite and identity of slopes in Figure 16 indicate that zircons can be used to typify granitic intrusives, confirming on a more rigid statistical basis statements by Groves (1930, p. 235) and Tyler *et al.* (1940, p. 1521–1522).

Figures 12, 13, 14, and 16 imply also that zircons in Bald Mountain tonalite and Anthony Lake granodiorite are of early crystallization. As crystal habit, shape, and size of any mineral is dependent on physico-chemical environment, evidence for late zircon crystallization. During crystallization of a magma, growing crystals of the essential minerals would tend to push aside pre-existing zircons and thus confine them to grain boundaries or areas of late crystalliza-

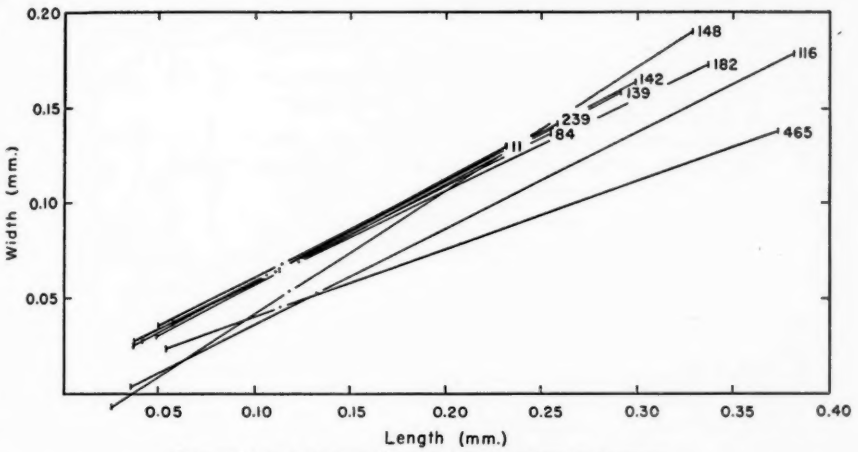

FIGURE 16.—REDUCED MAJOR AXES OF ZIRCON SAMPLES

Specimen number of each sample is at right end of corresponding reduced major axis. Length of line is determined by longest and shortest crystals of sample. Dots are position of mean length and width of respective samples.

statistical uniformity reflects a uniform environment. A uniform environment must be an environment that has not experienced the differentiation resulting from extensive crystallization. This uniform environment, reflected by the demonstrated uniformity of zircons throughout 167 square miles of tonalite and granodiorite, must have been the early environment that existed at the onset of crystallization before the major intrusive mass differentiated into a marginal tonalite that grades into a core of granodiorite. Statistical uniformity of zircons throughout a large intrusive that ultimately differentiated into tonalite and granodiorite is probably the most convincing evidence that these zircons crystallized early.

The conclusion that zircon in felsic igneous rocks is of early crystallization is contrary to statements by Moorhouse (1953, p. 1455) that zircon is a late (deuteric) crystallizer. Apparently Moorhouse bases his conclusion on the common occurrence of zircon along grain boundaries and the association of much zircon with the essential minerals of late crystallization. However, these relations cannot be used as tion. Actually, such a process has not been very effective, as numerous zircons in the Bald Mountain tonalite and Anthony Lake granodiorite occur within the early crystallizing plagioclase and hornblende. More significant, the statistical uniformity expressed by Figures 12, 13, 14, and 16 is stronger evidence for early crystallization.

Prior studies of felsic igneous zircons have shown that peaks for length-breadth curves (Smithson, 1939, p. 351) and elongation-frequency curves (Coetzee, 1942, p. 100; Poldervaart and von Backstrom, 1949, p. 475) reach maxima at ratios greater than 2. These results apply only to zircons of some granitic rocks; Figure 14 shows that zircons of tonalites and granodiorites may have elongation frequency ratios having maxima less than 2.

CHEMICAL DATA

Chemical analyses and norms of 12 rocks are given in Table 18. Modes of the analyzed specimens appear in Table 19. Although the Willow Lake norite and Badger Butte quartz-

TABLE 18.—CHEMICAL ANALYSES AND NORMS

H. B. Wiik, Helsinki, Finland, Analyst

	A	B	C	D	E	F	G	H	I	J	K	L
SiO_2	51.83	56.60	58.82	66.18	67.53	72.40	72.43	73.04	73.96	76.55	76.47	76.68
TiO_2	0.77	0.72	0.57	0.50	0.46	0.21	0.17	0.20	0.16	0.10	0.15	0.09
Al_2O_3	21.45	18.67	15.98	16.58	15.74	15.67	14.44	15.01	13.91	12.70	12.60	12.88
Fe_2O_3	0.90	1.14	0.25	1.40	1.24	1.28	1.03	0.74	1.03	0.39	0.59	0.88
FeO	5.90	5.25	6.26	2.59	2.56	1.15	1.05	0.79	0.79	0.72	0.13	0.13
MnO	0.18	0.17	0.19	0.09	0.10	0.04	0.02	0.04	0.04	0.04	0.00	0.00
MgO	4.68	4.60	5.83	1.81	2.16	0.27	0.92	0.75	0.25	0.19	0.50	0.25
CaO	8.88	7.58	8.39	5.09	3.89	2.89	2.52	3.36	2.10	1.07	0.34	0.79
Na_2O	3.97	3.40	2.15	3.76	3.57	3.71	4.11	3.86	3.32	3.22	2.36	3.66
K_2O	0.34	0.95	0.72	1.19	2.14	1.93	2.82	2.35	3.73	4.36	6.31	5.01
P_2O_5	0.33	0.29	0.08	0.15	0.13	0.08	0.11	0.10	0.06	0.02	0.07	0.00
H_2O+	1.08	0.80	0.94	0.43	0.62	0.31	0.84	0.30	0.31	0.22	0.84	0.02
H_2O-	0.06	0.09	0.03	0.07	0.11	0.00	0.10	0.08	0.03	0.01	0.17	0.05
	100.37	100.30	100.21	99.84	100.25	99.94	100.56	100.62	99.69	99.59	100.53	100.44
					Norms							
qu	.....	7.62	14.04	24.96	25.74	36.54	31.02	33.48	35.10	38.16	37.37	34.32
or	2.00	6.12	4.45	7.23	12.78	11.68	16.68	13.90	22.24	25.80	37.25	29.47
ab	33.54	28.82	18.34	31.96	30.39	31.44	34.58	32.49	28.30	27.25	19.91	30.92
an	39.48	32.53	31.69	24.46	18.34	13.62	11.68	15.85	10.29	5.31	1.29	3.89
cor	.....	.....	.....	.10	.61	2.45	.31	.31	.51	.71	1.42	.....
di	2.29	2.72	7.51	.....	.....	.....	.....	.....	.....	.....	.....	.....
hy	14.77	17.96	21.61	7.40	8.57	1.50	3.22	2.43	1.13	1.43	1.25	.60
ol	3.59	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
il	1.52	1.37	1.06	.91	.91	.46	.30	.46	.30	.15	.30	.15
mt	1.39	1.62	.46	2.09	1.62	1.86	1.39	1.16	1.39	.56	.....	.19
hm	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.59	.75
ap	.77	.67	.33	.33	.33	.33	.33	.33	.....	.....	.17	.....
H_2O	1.14	.89	.97	.50	.73	.31	.94	.38	.34	.23	1.01	.07
	100.49	100.32	100.46	99.94	100.02	100.19	100.45	100.79	99.58	99.60	100.56	100.36

Table 18 (*Continued*)

A (Spec. 539)	Willow Lake norite, 1000 feet southeast of Willow Lake.	G (Spec. 447)	Granodiorite porphyry dike rock, 1 mile north of top of Red Mountain.
B (Spec. 521)	Badger Butte quartz-biotite norite, 800 feet northeast of Badger Butte.	H (Spec. 471)	Red Mountain leucogranodiorite, 1.3 miles northeast of top of Red Mountain.
C (Spec. 395)	Black Bear quartz gabbro, 2 miles southeast of top of Hunt Mountain.	I (Spec. 466)	Elk Peak leucocratic quartz monzonite, 1000 feet southeast of top of Elk Peak.
D (Spec. 266)	Bald Mountain tonalite, 3000 feet west of Baldy Lake.	J (Spec. 491)	Leucocratic quartz monzonite dike rock, 1 mile northwest of Red Mountain Lake.
E (Spec. 145)	Anthony Lake granodiorite, 1.5 miles north of Anthony Lake.	K (Spec. 136)	Rhyolite porphyry dike rock, 0.9 mile northeast of Red Mountain Lake.
F (Spec. 152)	Mount Ruth leucogranodiorite, 300 feet north of top of Mount Ruth.	L (Spec. 98)	Aplite, 1 mile northeast of Anthony Lake.

biotite norite are mineralogically norites, chemical data show that these rocks are closer to diorites (Nockolds, 1954, p. 1019–1020).

The chemical analyses indicate that the alkali-lime index (Peacock, 1931, p. 56) is about 66, compared to an index of about 65 for rocks of the southern California batholith. The analyses and norms of the Bald Mountain tonalite and Anthony Lake granodiorite are compared in Table 20 to Nockolds' (1954, p. 1014–1015) average tonalite and granodiorite. A close similarity exists between the tonalites. The granodiorites show a slightly greater contrast because the Anthony Lake granodiorite is a potassium-poor rock transitional to a tonalite.

In Figure 17, curves for the rocks of the southern California batholith (Larsen, 1948, p. 142) are shown on a variation diagram of the type used by Larsen (1938, p. 506–507). Analyses of rocks from the Bald Mountain batholith have been plotted on Figure 17, and the position of the plots can be compared with Larsen's curves for the southern California batholith. The rocks of the Bald Mountain batholith are somewhat lower in FeO and K_2O, slightly higher in SiO_2, and noticeably higher in Al_2O_3. The Al_2O_3 and FeO plots for the Oregon rocks more nearly coincide to the curves shown by Larsen (1948, p. 145) for the plutonic rocks east of the Peninsular Ranges.

Nockolds and Allen (1953, p. 106) used a variation diagram showing the relations between Mg, $(Fe^2 + Fe^3)$, and (Na + K) and between Ca, Na, and K. Figure 18 is a reproduction of the curves shown by Nockolds and Allen (1953, p. 107) for the rocks of the southern California batholith and for the rocks of the Lassen Peak area. The rocks of the Bald Mountain batholith have been plotted on Figure 18 and fall near the variation curves for the California rocks.

The chemical data indicate that the rocks of the Bald Mountain batholith are similar to other calc-alkaline associations.

Evidence for Magma

General Evidence

The composite character of the batholith, with the various distinctive units emplaced in

TABLE 19.—MODES OF CHEMICALLY ANALYSED SPECIMENS

	A	B	C	D	E	F	G	H	I	J	K	L
Quartz	..	6.7	16.7	27.2	26.4	35.8	30.1	34.0	36.9	38.4	36.4	38.6
Potassium Feldspar	..	..	..	1.4	7.2	9.5	18.0	9.2	26.7	30.0	33.8	32.5
Plagioclase	65.6	57.7	47.6	53.0	53.8	50.5	42.9	53.5	34.2	28.9	20.7	28.4
Biotite	..	9.0	4.9	10.5	7.0	4.0	..	3.1	1.9	2.2	..	0.2
Hornblende	12.8	13.9	7.2	6.0	4.1	..	..	..	..	..	..	..
Pyroxenes	3.1	1.9	2.1	..	..	..	..	..	..		..	..
Actinolite	15.5	8.8	20.0	..	..	..	..	..	..	..	..	
Chlorite	..	..	..	..	..	..	6.2	..	..	..	3.6	..
Epidote	..	..	..	..	..	..	2.4	..	..	..	..	..
Sericite	..	..	..	..	..	..	..	..	..	..	5.1	..
Opaque ores	2.6	1.5	1.2	1.6	1.1	0.2	0.2	0.2	0.2	0.4	0.3	0.3
Nonopaque accessories	0.4	0.5	0.3	0.3	0.3	..	0.2	..	0.1	0.1	0.1	..

TABLE 20.—CHEMICAL ANALYSES AND NORMS OF AVERAGE TONALITE AND GRANODIORITE COMPARED WITH BALD MOUNTAIN TONALITE AND ANTHONY LAKE GRANODIORITE

	Bald Mountain tonalite	Average tonalite	Anthony Lake granodiorite	Average granodiorite
SiO_2	66.34	66.15	67.43	66.88
TiO_2	0.50	0.62	0.46	0.57
Al_2O_3	16.62	15.56	15.72	15.66
Fe_2O_3	1.40	1.36	1.24	1.33
FeO	2.60	3.42	2.56	2.59
MnO	0.09	0.08	0.10	0.07
MgO	1.81	1.94	2.16	1.57
CaO	5.10	4.65	3.88	3.56
Na_2O	3.77	3.90	3.56	3.84
K_2O	1.19	1.42	2.14	3.07
H_2O	0.43	0.69	0.62	0.65
P_2O_5	0.15	0.21	0.13	0.21
Norms				
qu	25.1	24.1	25.7	21.9
or	7.2	8.3	12.8	18.3
ab	32.0	33.0	30.4	32.5
an	24.5	20.8	18.3	16.4
cor	0.1	...	0.6	...
$CaSiO_3$	...	0.3	...	...
$MgSiO_3$	4.5	4.9	5.4	3.9
$FeSiO_3$	2.9	4.1	3.2	2.9
mt	2.1	2.1	1.6	1.9
il	0.9	1.2	0.9	1.1
ap	0.3	0.5	0.3	0.5
No. of analyses	(1)	(58)	(1)	(137)

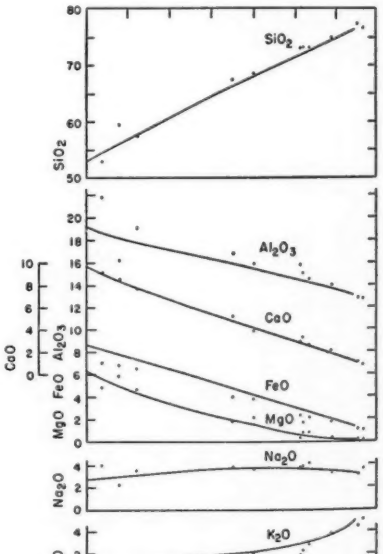

FIGURE 17.—VARIATION DIAGRAM OF ROCKS OF THE BATHOLITH OF SOUTHERN CALIFORNIA AND PLOTTED ANALYSES OF BALD MOUNTAIN BATHOLITH

Curves are for batholith of southern California and dots for Bald Mountain batholith.

a sequence of increasing silica content, lends support to the concept of magmatic origin. The presence of fractured remnants of metamorphosed screens between some members of

the intrusive sequence is also in keeping with the emplacement of discrete masses of magma.

In the Elkhorn Mountains the most extensive and best exposures of pre-batholithic rocks are on the southeast side of the batholith in a rugged area of about 80 square miles. Regional metamorphism is of low grade. The sediments belong to the greenschist facies. This low-grade metamorphic environment is entirely out of harmony with metamorphic intensities necessary for granitization. Read (1949, p. 143–151) similarly concluded that granitic plutons are of magmatic origin if they occur in areas of low-grade regional metamorphism.

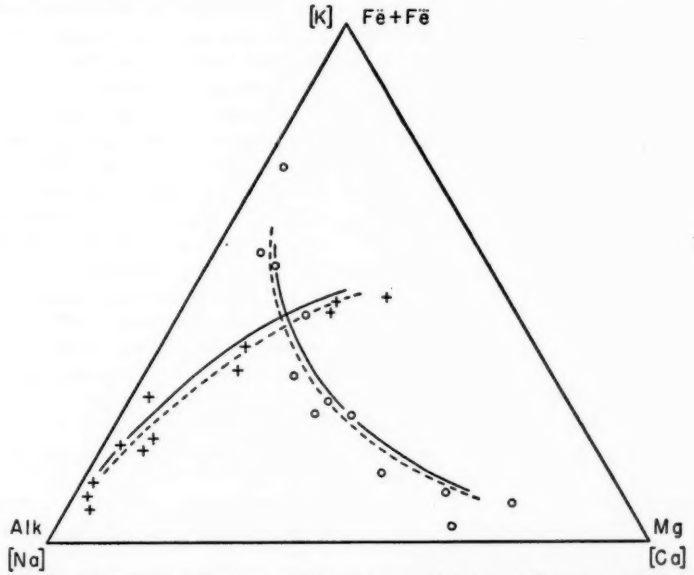

FIGURE 18.—VARIATION DIAGRAMS OF ROCKS OF THE SOUTHERN CALIFORNIA BATHOLITH AND THE LASSEN PEAK AREA COMPARED WITH ANALYSES OF ROCKS OF BALD MOUNTAIN BATHOLITH

Solid curves for rocks of southern California batholith are after Figure 1, and dashed curves for rocks of Lassen Peak area are after Figure 2 of Nockolds and Allen (1953). Crosses refer to variations in (Fe^2 + Fe^3), Mg, and (Na + K) for rocks of Bald Mountain batholith; circles refer to variations in Ca, Na, and K.

Bald Mountain Tonalite and Anthony Lake Granodiorite

The features of granite tectonics seem best interpreted as indicating the former existence of a magma. Likewise, the disrupted regional trend suggests the addition of enough introduced material to warrant the assumption that the introduced material had the characteristics of magma. Additions and substractions during granitization might cause a slight disturbance of the regional trend, but a static process such as granitization could not produce the marked disruption of strata that occurs around the northeastern Oregon plutons. Although the origin of the distributive faulting is uncertain, this widespread faulting can be more readily related to the emplacement of a magma than to granitization of country rocks *in situ*.

Diversely oriented inclusions near batholithic contacts testify to movement. Excellent examples of schist fragments showing movement occur in many places along the southern contact where apophyses and sharply cross-cutting dikes also contain numerous inclusions of random orientation. One large hand specimen from a dike contains three schist inclusions,

each of which has a different attitude from that of its neighbors or that of the wall rock. Along some of the brecciated batholithic contacts between Big Cracker Creek and the Baisley-Elkhorn mine, angular fragments of country rock can be fitted together, and in several localities small fragments can be fitted back into the wall rock. All these features are evidence for the previous existence of a magma.

Sharp contacts are bordered by aureoles of thermally metamorphosed rock without intervening areas of granitization. This fact suggests the original presence of magma. The most convincing evidence for granitization of batholithic dimensions is obtained from places that contain schist-gneiss-migmatite-granite transitions on a regional scale. No such transitions exist around the Bald Mountain batholith. No gneiss or migmatite can be found, and the schists are restricted to marginal areas or to screens and wedges of intensely deformed country rock which project into the batholith. Aureoles of thermally metamorphosed rock such as those which surround the northeastern Oregon intrusives are characteristic of granitic masses having an orthomagmatic origin (Read, 1949, p. 143–151).

The general presence of sharp contacts is regarded as evidence favoring magmatic intrusion. Although the significance of sharp contacts has been questioned (Reynolds, 1947, p. 216), nevertheless it is difficult to visualize a wave of static metasomatism on a front of many miles abruptly halting to produce contacts such as those which surround the Bald Mountain batholith.

Petrographic evidence confirms the conclusion that the Bald Mountain tonalite and Anthony Lake granodiorite are of magmatic derivation.

The remarkably small variation in the composition of the Bald Mountain tonalite and the Anthony Lake granodiorite strongly suggests the existence of a homogeneous parent material. These two very closely related rocks, representing about 97 per cent of the exposed intrusive complex, exist in a region characterized by marked petrographic heterogeneity. The invaded terrain is composed of metagabbro, serpentinized ultramafic rocks, and the Elkhorn Ridge argillite series that includes limestone lenses, greenstones, and a large percentage of chert. In fact, it is estimated that the Elkhorn Ridge argillite contains nearly 50 per cent of chert and cherty argillite. Point-counter analyses indicate that the granitic intrusives have a composition which is indifferent to the areal distribution of the heterogeneous assortment of country rocks. Intrusive contacts that pass from recrystallized chert into serpentinized ultramafic rocks show little or no change in the composition of the adjacent tonalite. It seems impossible that any process of static metasomatism could transform these chemically diverse pre-batholithic rocks into granitic rocks which display such a remarkably small variation in composition, as do the Bald Mountain tonalite and the Anthony Lake granodiorite. Where granitization has occurred in this chemically contrasting northeastern Oregon terrain, as in the area southwest of Badger Butte, the granitized rocks have no uniformity of composition or of texture and are characterized by a striking variability in thin sections as well as in field outcrops. Such granitized rocks are so distinctive that there can be no confusion with rock types within the batholith. The petrographic uniformity of the batholithic rocks is in harmony with the concept of a magmatic origin.

Gorai (1951, p. 900) suggested that twinning characteristics of the plagioclase can be used to distinguish magmatic granites from metasomatic granites. In a detailed investigation Gorai (1951, p. 890–898) used a U-A-C diagram to illustrate the contrast in the plagioclase twinning of igneous and metamorphic rocks. Figure 19 is a U–A–C. diagram that shows the character of the plagioclase twinning in the various rocks of the Bald Mountain batholith. All the plutonic rocks in the Bald Mountain batholith fall within the igneous field of Gorai, and the relative position of each corresponds with the results obtained by Gorai (1951, p. 892) from his study of igneous rocks in the gabbro to granite range.

The uniformity in composition of the plagioclase cores supports the concept that they crystallized from an essentially homogeneous magma. Examination of plagioclase from widely separated samples of Bald Mountain tonalite and Anthony Lake granodiorite

shows that cores of these crystals fall in a restricted range between An_{48-44}. As the plagioclase cores represent the first salic material to crystallize, it seems logical that it crystallized from a more or less homogeneous magma rather than that these nearly uniform plagioclase cores grew in an essentially solid and contrasting environment of serpentine, metagabbro, chert, limestone, and argillite. Moreover, the zoned structure of the plagioclase is comparable to zoning in plagioclase of undoubtedly magmatic rocks. Plagioclase phenocrysts in many rhyolites, dacites, and acid andesites are practically identical in their zoning characters to the plagioclase in the batholithic tonalite and granodiorite. Assuming that this identity can be attributed to a similarity in environment, it follows that the batholithic plagioclase also passed through a silicate melt stage. This seems the most likely possibility, in view of the lack of similar zoning in most rocks of unquestioned metamorphic origin.

The uniformity in composition of hornblende in the Bald Mountain tonalite and Anthony Lake granodiorite implies that this mineral crystallized from an essentially homogeneous magma. The consistency of composition can be established by the uniformity of 2V and γ index. Disregarding specimens less than 200 yards from the contact, optic axial-angle determinations on 79 different thin sections from an area representing approximately 160 square miles of rock gave a range in $2V_x$ from 65.5° to 72°. Hornblende was the first mafic mineral to crystallize, and the small range in 2V supports the concept of a magmatic origin. Determinations of $2V_x$ on 17 different specimens from within 200 yards of the contact gave a range of 70°–79°. Apparently the character of the wall rock had no influence on optic axial angles, because 2V values for hornblende in tonalites adjacent to chert are similar to those adjacent to metagabbro. This lack of correlation between wall rock and hornblende composition also suggests a magmatic origin.

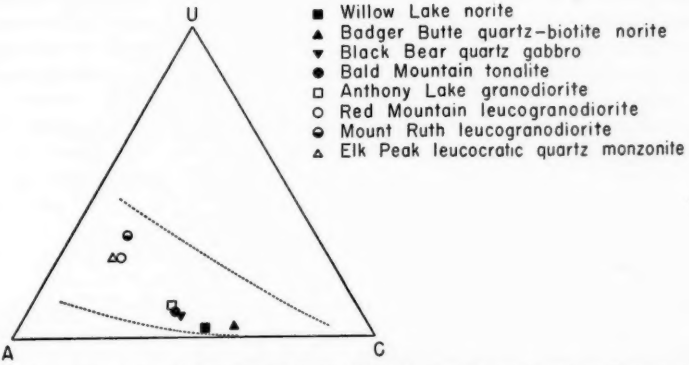

FIGURE 19.—DIAGRAM OF TWINNING HABITS (U-A-C) FOR PLAGIOCLASE IN ROCKS OF THE BALD MOUNTAIN BATHOLITH
Area between dashed curves is igneous field of Gorai (1951).

The hypidiomorphic granular texture of the tonalite and granodiorite suggests a magmatic origin; a crystalloblastic texture would imply a metamorphic origin. The batholithic hornblendes are commonly euhedral or subhedral in contrast to the ragged and unterminated hornblendes of many metamorphic rocks. The plagioclase of metamorphic rocks is typically xenoblastic in outline, whereas plagioclase of the tonalite and granodiorite has a distinct idiomorphic tendency. Where biotite is in contact with quartz and potassium feldspar, it commonly contributes to the hypidiomorphic texture, and some quartz shows crystal faces against potassium feldspar. These features seem best reconciled as having originated in a crystallizing magma and not in a metamorphic

environment where each individual crystal in an essentially solid medium simultaneously exerted its own crystallization force against the resistance offered by competing adjacent crystals.

Zircon characters favor a magmatic origin for the tonalite and granodiorite. Poldervaart and Eckelmann (1955, p. 947) indicate that high proportions of zircons showing evidence of new growth are characteristic of autochthonous granite. Zircons from the batholith show no evidence of new growth. Zircons of the Bald Mountain tonalite and Anthony Lake granodiorite are entirely different from the sedimentary zircons of the metamorphic aureole (unpublished data). These facts support a magmatic origin for the tonalite and granodiorite.

Protoclastic zones along the outer margins of the tonalite imply the original presence of a magma. Although milder fracturing can be distinguished only in thin sections, crushing and shearing in some areas has been severe enough to be noticeable in the field. Under extreme conditions, a marked foliation is present, and a mylonite of plagioclase and hornblende crystals exists in a matrix of quartz and biotite of later crystallization. This crushing cannot be attributed to dynamic metamorphism following the consolidation of the intrusives. Marginal crushing of the tonalite can only be attributed to protoclastic effects during the consolidation of the magma.

Norites

A magmatic origin for the tonalite and granodiorite does not necessarily imply that the associated norites and the Black Bear quartz gabbro, grouped with norites for convenience, are also magmatic. All the norites are marginal to the batholith, and they might be former sediments converted to noritic rocks during the metamorphism that accompanied the emplacement of the tonalite and granodiorite. This possibility is rejected for the following reasons.

Although pyroxene-bearing rocks peripheral to the batholith denote the most intense areas of contact metamorphism, such rocks are limited to distances of not more than about 120 feet from the contact. Hence these pyroxene rocks crop out in a narrow and elongated strip, whereas the noritic bodies have the oval form of stocklike intrusives. In northeastern Oregon the difficulty of producing an essentially uniform rock having the form of an oval intrusive by static noritization of heterogeneous country rock seems almost insurmountable.

An abrupt contrast exists between the norites and the adjacent country rocks. Sharp contacts separate the medium-grained norites from the surrounding pyroxene hornfels characterized by a sugary texture and variable compositions. The lath-shaped plagioclase of the norites has igneous twinning (Fig. 19), whereas most of the anhedral plagioclase of the hornfels is untwinned.

Inclusions in the norites show evidence of movement. Excellent examples, some on a hand-specimen scale, of diversely oriented fragments of country rocks are present in the Black Bear intrusive. Regardless of the origin of the remarkable banding in the Willow Lake norite, the presence of broken and dismembered pieces of this banded rock scattered throughout certain parts of the norite indicates the former existence of a noritic magma.

Along part of their borders, all the norites are separated from the Bald Mountain tonalite by screens of country rock. These screens are not of a uniform rock type which might be postulated as resisting possible noritization processes. The screens contain every type of country rock except serpentine. The metamorphosed limestones, greenstones, argillites, siliceous argillites, cherts, and amphibolites of the screens do not resemble the adjacent norites. This lack of resemblance suggests that the screen rocks and norites are of different origin. The screen rocks are shattered and fragmented, but they tend to box the compass around the norites. This feature is considered a result of a shouldering aside of the country rocks by an intrusive magma (Noble, 1952, p. 54).

The Willow Lake and Badger Butte norites possess a foliation. The foliation has not been mapped in detail except in the Willow Lake norite, and this mapping has not been completed. However, the unfinished structural examination indicates a foliation pattern in the

Willow Lake norite that can only be attributed to intrusive movements involving magma.

The sharp contrast that exists between the zircons from the metamorphosed sediments (unpublished data) and the zircons from the norites does not support the concept of a metasomatic origin for the norites. As it is almost inconceivable that the features of the noritic zircons could have survived sedimentary processes, these zircons seem to indicate a magmatic origin for the norites.

Leucocratic Granodiorite and Quartz Monzonite

The following factors were instrumental in concluding that the leucocratic units of the batholith have crystallized from discrete masses of magma and have not originated in any other manner.

All the leucocratic rocks have sharp contacts with the tonalite. These sharp contacts are considered evidence for the intrusion of magmas. Brecciated pieces of tonalite occur along contacts of the Red Mountain leucogranodiorite and along contacts of the Elk Peak leucocratic quartz monzonite. These fragments of tonalite imply forceful emplacement.

The outcrop pattern of the leucocratic rocks is difficult to visualize as originating through any process other than intrusion of magmas. Apophyses and small dikes extend from the Red Mountain and Elk Peak rocks into the adjacent tonalite. These relations indicate the former presence of magma. Although no leucogranodiorite dikes can be traced into the Mount Ruth rock, dikes of leucogranodiorite near Mount Ruth probably are connected with this intrusion.

The intrusive sequence of the leucocratic rocks indicates the emplacement of separate magmas. The Elk Peak quartz monzonite cuts a small mass of Red Mountain leucogranodiorite, and a dike having the composition of the Elk Peak quartz monzonite cuts the Mount Ruth leucogranodiorite. Not only do these relations indicate the existence of separate magmas, but the fact that the younger rock is always more silicic suggests that some systematic magmatic process such as fractional crystallization may have operated to control the emplacement sequence.

Zircons of the leucocratic rocks are different from those found in the metamorphic aureole, and features that contribute to this pronounced difference are interpreted as indicating that the zircons of the leucocratic rocks are magmatic. Zircons of each leucocratic rock are also distinct from other batholithic zircons and are therefore regarded as originating in different physico-chemical environments. As the zircons of the leucocratic rocks are considered of magmatic origin, the existence of these different environments indicates the former presence of different magmas.

Method of Emplacement

General Statement

In view of the controversy of recent years regarding emplacement mechanisms for granitic masses of batholithic dimensions, one of the major considerations of this investigation was to observe features that might reveal processes of emplacement. As field observations are of paramount importance (Read, 1946, p. 666; 1948, p. 7), the emphasis was placed upon detecting features which exist in the field.

Emplacement of granitic plutons is generally acknowledged to be the result of one or combinations of the following three major processes:

(1) Forceful intrusion
(2) Piecemeal stoping with or without major foundering
(3) Granitization

In this paper granitization is used to denote the formation of granitic rocks by static metasomatic transformations of rocks which previously were less granitic in appearance and composition. Melting in place with essentially no subsequent movement before consolidation might be another possibility for the emplacement of granitic plutons.

The Bald Mountain batholith was emplaced as a result of a combination of the three major processes listed. The relative importance of each of the three processes is not known, but the approximate importance of each seems well established. Apparently the dominant emplacement factor was forceful injection. Stoping also contributed to emplacement, but the magnitude of the stoping process is unknown. Locally granitization occurred, but regionally it was of little consequence. It is

uncertain how much importance should be attached to the stoping process. Evidence for piecemeal stoping is conclusive, but in exposures now visible the space acquired in this manner represents only a minor fraction of the total volume occupied by batholithic rocks. Although no substantiating evidence is preserved in the outcrops, it is conceivable that stoping was of greater importance and may have operated on a larger scale during the early stages of intrusion. Field observations support the conclusion that stoping should be given a position of secondary rank and that forceful injection was the major factor responsible for the emplacement of the Bald Mountain batholith and its associated stocks.

Evidence for Forceful Injection

The strong deflection of the regional trend around the intrusives suggests that the country rocks have undergone a three-dimensional adjustment to accommodate the magmatic intrusions. There are two possible methods by which structural trends in the country rock could be wrapped around a granitic pluton. An intrusive magma might forcefully shoulder aside its walls, or forces of external origin might mold the wall rocks to the shape of a solid intrusion. In northeastern Oregon the latter possibility is not applicable. There is no regional cataclastic crushing of the intrusive margins, such as that which should occur if the country rock had been molded to the shape of a solid intrusion. Dikes extending outward from the intrusives are not shattered and dislocated, as they would be if they had been subjected to external forces capable of disrupting regional trends to fit the configuration of intrusives. Shattering and crushing might not occur under conditions of plastic deformation, but Griggs (1936, p. 550–554) has shown that even under plastic conditions both fracture and flow can occur simultaneously. Moreover, in this part of northeastern Oregon, every deformation since the original intense folding of the Elkhorn Ridge argillite has been characterized by extensive fracturing, and plastic phenomena have been at a minimum. Since the batholithic emplacement, the surrounding area has behaved as a rigid block. Most later Cretaceous deformation was confined to the formation of regional shears trending east-west to northeast, which cut all rocks indifferently. These shear zones, some of which can be followed for miles, have not been affected appreciably by Tertiary deformation. Tertiary deformation consisted of a breaking into smaller units which moved differentially in conjunction with mild folding of the late Cretaceous gravels (Taubeneck, 1955b, p. 95) and Tertiary volcanics. These gravels and volcanics rarely have dips greater than 20°. Hence there is no evidence that subsequent forces were responsible for molding the structural trends to the form of the intrusives. The remaining alternative is that the country rocks box the compass around the intrusives, because forcefully injected magmas pushed and shouldered aside the retaining walls.

The distributive faulting is found only in pre-batholithic rocks. The pattern of this faulting cannot be reconciled with the directional forces responsible for the tight folding of the Elkhorn Ridge argillite. It seems most reasonable to refer the origin of these faults to the forceful injection of the granitic masses. These innumerable distributive faults are regarded as part of the adjustments which took place in the country rocks, in order to accommodate the intruding magmas.

In many places there are breccias along the intrusive contacts. Shattered contacts along the margins of the batholith and similar features between the component members of the intrusive sequence attest to an origin by forceful emplacement.

The structural features of granite tectonics indicate forceful intrusion. The presence of marginal joints and thrusts supports the concept of a strong upward movement in the main plutonic body. Such joints and shears seem best interpreted as originating from intrusive pressure forcing the wall rock both upward and outward. The linear flow structure and the foliated borders with platy parallelism reveal the resistance to motion of the moving crystal mush past confining wall rocks. Another notable feature of granite tectonics is the elongated and spindle-shaped inclusions that parallel the intrusive contacts. The intense stretching and drawing out of these inclusions testifies to

the forceful emplacement of the Bald Mountain batholith.

Evidence for Stoping

Field evidence for stoping is not so widespread or so impressive as the evidence for forceful injection. Nevertheless, enough evidence exists to substantiate the conclusion that stoping played at least a subordinate role in the emplacement processes.

The upper part of cupolalike extensions of granitic rock contain numerous inclusions of random orientation that have been stoped from overlying country rocks having a uniform attitude. An excellent illustration of this process can be seen approximately 1 mile S. 45° W. from the summit of Hunt Mountain. At this locality a precipitous stream has cut down through the overlying country rocks to reveal the top of a small cupola. A sharp intrusive contact is present, and within the tonalite there are stoped blocks with greatest dimensions of about 10 feet. These inclusions represent between 5 and 10 per cent of the total exposed area of the cupola. They are well defined, and many of them show a marked difference in orientation from that of the overlying country rocks. Thus these fragments and blocks are stoped portions of the country rock which were incorporated within the tonalite during the later stages of its cooling history.

Along some of the brecciated batholithic contacts the angular fragments can be fitted together, and in several places they can be fitted back into the wall rock. This is a small-scale phenomenon involving fragments measured in inches or a few feet.

Petrogenesis

Petrographic and field evidence indicates that all units of the Bald Mountain batholith are of magmatic derivation. Each unit represents a discrete volume of magma that has been forcefully emplaced in a regular sequence from norite to quartz monzonite. The various units might be differentiates of a parental magma of probable noritic[1] composition, or they could be derivatives of different magmas.

Certain facts seem to support the concept that crystal fractionation of a noritic magma may have been the major process responsible for the progressive variation in the intrusive sequence. If the intrusives are arranged as nearly as possible according to their order of emplacement, minerals in the intrusives appear and disappear in a definite order that coincides with Bowen's reaction series. The systematic change in mineralogy follows the course of variation that could result from fractional crystallization, and the close correlation between order of intrusion and order of crystallization suggests that this series of rocks formed from a mafic magma by processes involving crystal-liquid equilibrium. The parallelism between the intrusive sequence and crystallization sequence is shown in Table 21.

If the rock sequence in the batholith owes its origin to processes involving crystal-liquid equilibrium, the final crystallization products should approach or fall on the plagioclase-potassium feldspar-quartz cotectic curve (Nockolds, 1947, p. 19–28). The aplite is very near the plagioclase-potassium feldspar-quartz cotectic curve (Fig. 20). This is consistent with the viewpoint that aplites represent residual liquids derived by crystallization-differentiation from a more mafic magma. In Figure 21, the 12 chemically analysed rocks have been plotted on a potassium feldspar-albite-anorthite triangular diagram, and the position of the respective points can be compared with the points and crystallization curve of some Caledonian rocks selected by Nockolds (1946, p. 213). Disregarding the metamorphosed Black Bear quartz gabbro, a smooth crystallization curve could be drawn through points representing the Oregon rocks. This speculative curve would closely parallel the curve of Nockolds (1946, p. 215). As the Oregon plots, which represent a series of rocks ranging from

[1] The Willow Lake norite and Black Bear quartz gabbro are the most logical choices for a parental magma. The Willow Lake norite is chemically intermediate beteen the average norite and diorite of Nockolds (1954, p. 1019–1020). Before metamorphism, the original Black Bear rock was more mafic, and even now the normative feldspar and hypersthene of the metamorphosed quartz gabbro are close to the average norite. By subtraction of SiO_2 and K_2O introduced during metamorphism, the chemical composition of the Black Bear rock would approach the average norite. Probably the Black Bear magma should be regarded as the most logical parental magma.

norite to quartz monzonite, would fall on or near a single hypothetical crystallization curve, this fact apparently supports the concept of fractional crystallization of a noritic magma.

lith was systematic. As the earlier intrusions were norites and subsequent intrusions regularly changed toward granite, the logical deduction might be that this mafic to felsic intrusive

TABLE 21.—PARALLELISM BETWEEN INTRUSIVE SEQUENCE AND CRYSTALLIZATION SEQUENCE

Rock type	Rhombic pyroxene	Augite	Hornblende	Biotite	Plagioclase	Quartz	K-feldspar	
Willow Lake norite	X	X	X	(X)	X	..	..	Earliest
Badger butte quartz-biotite norite	X	X	X	X	X	X	..	
Bald Mountain tonalite	..	..	X	X	X	X	X	
Mount Ruth leucogranodiorite	..	..	..	X	X	X	X	
Red Mountain leucogranodiorite	..	..	..	X	X	X	X	
Elk Peak leucocratic quartz monzonite	..	..	..	X	X	X	X	
Aplites	..	..	..	(X)	X	X	X	Latest

(X)—A minor constituent.

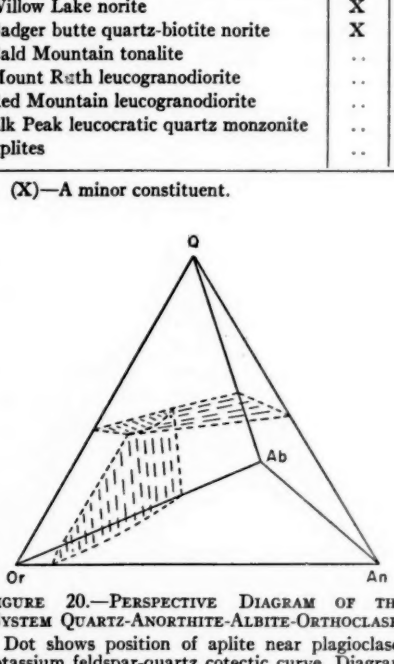

FIGURE 20.—PERSPECTIVE DIAGRAM OF THE SYSTEM QUARTZ-ANORTHITE-ALBITE-ORTHOCLASE

Dot shows position of aplite near plagioclase-potassium feldspar-quartz cotectic curve. Diagram after Nockolds (1946, p. 20).

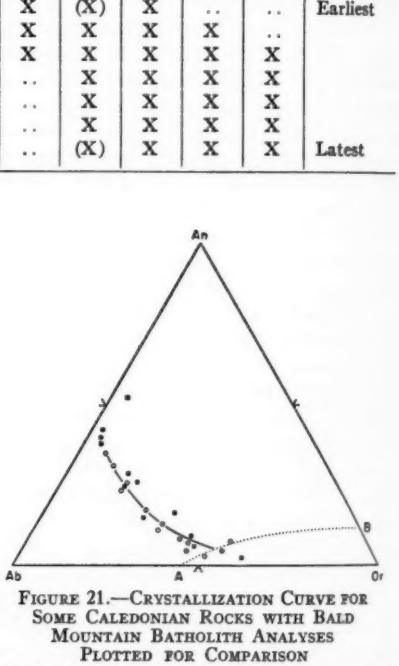

FIGURE 21.—CRYSTALLIZATION CURVE FOR SOME CALEDONIAN ROCKS WITH BALD MOUNTAIN BATHOLITH ANALYSES PLOTTED FOR COMPARISON

Dashed curve and circles indicate a course of crystallization in the plagioclase field for some Caledonian plutonic rocks (after Nockolds, 1946). Rocks of Bald Mountain batholith, plotted as squares, show similarity to Caledonian rocks. Cotectic curve between potassium feldspar and plagioclase is indicated by dotted line AB.

The analyses of the batholithic rocks have been plotted on a variation diagram (Fig. 17) of the type used by Larsen (1938, p. 506–507), and the analyses fall near smooth variation curves such as those for the southern California batholith (Larsen, 1948, p. 142) which are typical in general shape to the curves of other calc-alkaline plutonic associations. The smooth variation curves which could be drawn for the Oregon rocks and the general similarity to curves of other calc-alkaline associations suggest that the dominant process that produced the rock variation in the Bald Mountain batho-

series was produced by crystal fractionation of a parental noritic magma.

Nockolds and Allen (1953, p. 106–115) investigated seven igneous rock series of the calc-alkaline type. They used a variation diagram showing the relations between Mg, (Fe^2 + Fe^3), and (Na + K) and between Ca, Na, and K to indicate similarities between these igeneous rock series. Regardless of

whether or not the series was volcanic or plutonic, each series revealed a similar variation, and the conclusion was that all these series were magmatic. In addition, the smooth curves were believed to represent the liquid line of descent of mafic parental magmas. Figure 18 is a reproduction of the variation diagrams shown by Nockolds and Allen (1953, p. 197) for the rocks of the southern California batholith and for the rocks of the Lassen Peak area. The rocks of the Bald Mountain batholith have been plotted on Figure 18, and these rocks fall near the variation curves for the California rocks. Nockolds and Allen's conclusions should thus apply to the rocks of the Bald Mountain batholith, and the composition of the supposed parental magma should be similar to that of the Willow Lake norite.

Nockolds (1940, p. 496–499) chose the pyroxene-mica-diorite as the parent magma for the Garabal Hill-Glen Fyne complex. Disregarding xenoliths and the mafic accumulative rocks, the analyses from the complex have been plotted in Figure 22 on a (Na_2O + K_2O), MgO, and FeO triangular diagram of the type used by Wager and Deer (1939, p. 314). The circles indicate the position of pyroxene-mica-diorites, and the solid curve shows the differentiation trend of the Scottish rocks. A similar curve could be drawn for the Oregon rocks, and by analogy the parental magma of the Bald Mountain batholith should have been of noritic composition.

Calc-alkaline associations have often been attributed to fractional crystallization of a mafic magma. Some evidence has been presented that might be utilized to support the contention that the plutonic rocks of the Bald Mountain batholith have formed from a parental magma of noritic composition. However, the validity of such a concept is questionable, as there are conspicuous features that apparently do not reflect a serial derivation from mafic magma.

The frequency distribution of noritic rocks is entirely out of proportion to the great quantities of mafic magma that would be necessary for the production of the intermediate batholithic rocks through the operation of fractional crystallization. Grout (1926, p. 549) calculated that a maximum of one-tenth of an average basaltic magma may become granite. Actually, differentiation of the Duluth gabbro produced only 2 per cent of residual granophyric magma (Grout, 1918, p. 639). Wager and Deer (1939, p. 293) indicate that about 1 per cent of acid

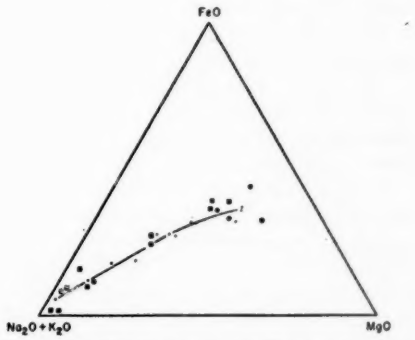

FIGURE 22.—DIFFERENTIATION TREND OF GARABAL HILL-GLEN FYNE COMPLEX WITH BALD MOUNTAIN BATHOLITH ANALYSES PLOTTED FOR COMPARISON

Curve and dots show differentiation trend of Garabal Hill-Glen Fyne complex. Four large circles are pyroxene-mica-diorites, regarded by Nockolds as the parental magma. For comparison, squares for rocks of Bald Mountain batholith might imply a parental magma of noritic composition.

differentiates resulted from strong fractionation of the Skaergaard magma. By careful selection, Nockolds (1940, p. 503) obtained 30 per cent of residual granodiorite from the fractional crystallization of a pyroxene-mica-diorite. These examples indicate that a given volume of mafic magma, even if contaminated by reaction with silicic material, can yield only a relatively small quantity of a more felsic differentiate. The norites constitute less than 3 per cent of the plutonic rocks in the batholith and stocks of the Elkhorn Mountains. In the Wallowa batholith (Taubeneck, in preparation) and other granitic intrusives of northeastern Oregon, norites are even less abundant. The great preponderance of tonalite and granodiorite in the northeastern Oregon intrusives strongly suggests that these rocks did not originate by simple or modified fractional crystallization of a noritic magma.

If the tonalite and granodiorite are differentiates of a mafic magma, volume requirements would necessitate that tremendous quantities

of mafic rock must underlie northeastern Oregon. It is difficult to understand how so much mafic material could now underlie northeastern Oregon and yet allow this mountainous region to stand so high isostatically.

An important consideration suggesting that fractional crystallization of a mafic magma is not an acceptable explanation for the origin of tonalities and granodiorites is the restriction of these rocks to continental areas and the absence of equivalent dacitic and rhyodacitic differentiates in the volcanic islands of the intra-Pacific region.

Another point worthy of consideration is that the differentiation trend of the calc-alkaline series from an assumed basic magma is contrary to the trend of fractional crystallization of basaltic magma, as shown by Wager and Deer (1939, p. 314), Walker and Poldervaart (1949, p. 656–662), Tilley (1950, p. 52–54), and Cornwall (1951, p. 165–170). These investigations show that, regardless of contrasting geologic environment, the major period of fractional crystallization of basaltic magma produces residual liquids that become increasingly enriched in iron. If this is the normal course in the fractional crystallization of basaltic magma, intermediate rocks of the basaltic series have no chemical or mineralogical resemblance to the positionally equivalent diorites and tonalites of the calc-alkaline series (Poldervaart and Elston, 1954, p. 159). However, Bowen (1947, p. 273) contends that examples of differentiation such as the above represent dry conditions and that basaltic magmas enclosed by more competent roofs retain sufficient volatile constituents to produce members of the calc-alkaline series.

In view of volume considerations and possible discrepancies in the trend of fractional crystallization, sequential differentiation of a noritic parental magma seems a rather tenuous explanation for the other intrusive units of the batholith. If the volumetrically insignificant norites were not present in the Bald Mountain batholith, there would be no reason for postulating that disproportionately larger bodies of tonalite and granodiorite were differentiates of small noritic masses. Mutual association is no infallible criterion of origin from a common source. For example, although the intimate association of Carboniferous coals and underclays is almost universal, Weller (1930, p. 123) has shown that the coals and underclays are not genetically related. Petrologic associations such as spilites and ultramafic rocks are commonly found in the same terrain, but their mutual occurrence appears to indicate a tectonic association rather than origin from a common source. Similarly, regional tectonics combined with anatexis in the depths of a geosynclinal belt could account for the norite-tonalite association. Admittedly there are objections and major uncertainties regarding the operation of a speculative process that would produce limited quantities of mafic magma and tremendous quantities of tonalitic magma. Seismic evidence indicates that linear and curvilinear orogenic structures are associated with complex reverse faults extending to general depths of 650 km (Benioff, 1949, p. 1837). These orogenic faults are earthquake loci along which great quantities of heat are generated during alternate to-and-fro bending hat accompanies repeated movements (Benioff, 1954, p. 399). Benioff (1954, p. 399) postulates that generation of heat produces molten rock which eventually appears as volcanic manifestations. Heat generated along these reverse faults might also explain the norite-tonalite association. If a rapid transfer of heat into the higher portions of the earth's crust occurred along the fault zone, the rising heat as it passed upward through relatively mafic material might produce minor quantities of noritic magma that could be intruded upward along zones of crustal weakness to form the small noritic masses. As heat continued to rise, it would encounter materials of lower melting range, and perhaps larger quantities of a tonalitic magma might be generated rather rapidly. Perhaps by this stage the heat would tend to be dispersed by the tonalitic magma which probably would move upward along the same zones of weakness that previously served as emplacement avenues for the noritic magma.

Regardless of the explanation for present uncertainties concerning the origin of calc-alkaline plutonic rocks, the available information seems to indicate that the norite-tonalite association of the Bald Mountain batholith is an occurrence resulting from the injection

FIGURE 23.—GEOLOGIC MAP OF PART OF ELKHORN MOUNTAINS

of magmas of different derivation into the same major zone of weakness.

Petrographic evidence indicates that fractional crystallization did control the mineralogy of the major intrusive unit of the batholith. After emplacement, this mass crystallized from the margins inward, producing by fractional crystallization a marginal tonalite that grades into a core of granodiorite.

Derivation of the small leucocratic intrusives of the batholith by differentiation of the abundant tonalite or granodiorite might appear more logical than derivation of the leucocratic rocks from a noritic parental magma. However, certain facts suggest that the leucocratic units did not originate through fractional crystallization of the tonalite or granodiorite. Plagioclase of Mount Ruth leucogranodiorite has cores that are as calcic as plagioclase of the tonalite or granodiorite. Differentiation of the tonalite or granodiorite sufficiently advanced to produce a leucocratic end product devoid of hornblende and deficient in biotite should show a corresponding change in plagioclase composition. Plagioclase composition of Mount Ruth leucogranodiorite does not agree with the concept of fractional crystallization of the tonalite or granodiorite.

Early crystallization of zircons in the tonalite and granodiorite is a basic conclusion. Figure 16 shows that zircons of each leucocratic intrusive differ and also are distinct from zircons of the tonalite and granodiorite. Figure 16 emphasizes that zircons from the tonalite and granodiorite are from the same magmatic intrusive and also that zircons of each leucocratic intrusive are from different magmatic bodies. The differences in zircons indicate that the leucocratic rocks did not differentiate from the tonalitic or granodioritic magma but crystallized from discrete masses of disconnected magmas.

In summation, a significant feature of the Bald Mountain batholith is an intrusive series of mafic to felsic rocks which apparently are not derivatives from the same magma. Nevertheless, if analyses of these rocks are plotted, curves can be drawn showing a variation similar to the trends found in other calc-alkaline associations. As the systematic variation of these trends correlates reasonably with physico-chemical predictions of what may happen during the crystallization of mafic magmas, variations of this type are generally regarded as evidence of derivation from one magma. Such a conclusion is unwarranted. A basic but often ignored fact is that a variation diagram shows only variation and never mode of formation.

Considerations expressed previously indicate that rocks of the batholith crystallized from separate masses of magma that were not differentiates of earlier intrusive units. The origin of these separate magmas is speculative. Regional tectonics combined with anatexis in the depths of a geosynclinal belt might account for different magmas, but until geologists possess a rational systematic hypothesis built on fractional melting, any explanation incorporating anatexis will contain unsupported and indefinite statements. A more complete explanation for the calc-alkaline association must wait until present and possibly future theories combine additional knowledge and perhaps new principles not yet applied to the interpretation of petrogenetic problems.

Conclusions

A review and consideration of the petrographic and field evidence lends support to the conclusion that the Bald Mountain batholith is of "magmatic" derivation, as this term is generally used, and was emplaced by the forceful ascent of discrete volumes of magma. Although stoping initially may have made important contributions to batholithic emplacement, most of the evidence now preserved in the rocks represents processes active during the later and final stages of the emplacement history, and this evidence indicates that stoping was subordinate.

Sequential differentiation through fractional crystallization of a noritic parental magma is considered an inadequate explanation for the characteristics of the more felsic intrusive units of the Bald Mountain batholith. The tonalite and granodiorite, leucogranodiorites, and leucocratic quartz monzonite are regarded as derivatives of distinct and disconnected magmas that crystallized individually. The intrusive sequence of the Bald Mountain batholith

seems best reconciled to the supposition of an initial generation and intrusion of noritic magma, followed by a later independent production and intrusion of much larger volumes of tonalitic magma. During crystallization of the tonalitic magma, differentiation after emplacement produced a marginal tonalite which grades into a core of granodiorite. Subsequently, minor quantities of different leucocratic magmas were intruded to terminate the intrusive sequence.

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THREE PRE-WISCONSIN GLACIAL STAGES IN THE ROCKY MOUNTAIN REGION

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ABSTRACT

Till, believed to be of pre-Wisconsin age, is widely distributed throughout the Rocky Mountains, in most places as small isolated patches. Some deposits lack morainal form and lie only on interstream divides; others, having a poorly preserved morainal form, overlap from the canyons onto the divides.

The deposits tend to be finer-grained, more compact, and more strongly jointed than younger Wisconsin deposits. Commonly they have a thick very strongly developed mature soil profile preserved on them.

In some places, the physical relations of the deposits to one intracanyon erosion surface and two precanyon erosion surfaces suggest that they represent three stages of glaciation. This suggestion is confirmed by stratigraphic evidence in the La Sal Mountains, Utah, and in Glacier National Park, Montana, where three tills, each capped by a thick and very strongly developed mature soil profile, are superposed.

The three stages may represent the Nebraskan, Kansan, and Illinoian stages of the Mississippi Valley, but no convincing evidence for such a correlation has been found.

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INTRODUCTION

Till, believed to be of pre-Wisconsin Pleistocene age, was first recognized in the Rocky Mountain region by Hills (1883), who noted deeply weathered precanyon glacial deposits on Tongue Mesa at the northern edge of the San Juan Mountains in Colorado. Similar deposits have since been reported at many other places throughout the region. Recently, exposures in which three pre-Wisconsin tills are superposed and separated by zones of deep weathering and soil development have been found by the writer in the La Sal Mountains, Utah, and along the east slope of the mountains in Glacier National Park, Montana.

The purpose of this paper is to describe the stratigraphic sections at these exposures and to focus attention on the many other localities where pre-Wisconsin till has been recognized.

Deposits regarded by the writer as of pre-Wisconsin age are those commonly correlated with the Buffalo stage of Wyoming (Blackwelder, 1915) or the Cerro stage of Colorado (Atwood and Mather, 1932). Deposits correlated with the Bull Lake stage of Wyoming (Blackwelder, 1915) or the Durango stage of Colorado (Atwood and Mather, 1932), though thought to be of pre-Wisconsin and possible Illinoian age by some workers, are believed by others, including the writer, to be of early Wisconsin age and are not considered here.

Distribution of Pre-Wisconsin Till

The distribution of pre-Wisconsin till as currently recognized in the Rocky Mountain region is shown in Figure 1, which has been drawn from sources cited in Table 1. The topographic setting of the several areas listed and the location of places referred to but not shown on Figure 1 may be found on the topographic or National Forest maps given for each area in Table 1.

Aspect and Character of Pre-Wisconsin Till

General Statement

Throughout the Rocky Mountain region pre-Wisconsin till tends to have a similar aspect and similar characteristics that are distinct from those of younger tills. The characteristics can be grouped into three categories: (1) the present physiographic habit of the deposits, (2) the physical character of the deposits at depth, and (3) the degree of weathering and soil-profile development on the deposits.

Physiographic Habit

Deposits of pre-Wisconsin till tend to be sheetlike in form and to lack morainal topography, though vestiges of moraines or kettles have been noted locally (Cross, Howe, and Irving, 1907, p. 7; Hole, 1910, p. 9; Pardee, 1925; Atwood and Mather, 1932, p. 102; Bradley, 1936, p. 194; Atwood and Atwood, 1938a, p. 239; Retzer, 1954, p. 28; Richmond, 1955, U. S. Geol. Survey open-file report; Ph.D. thesis, Univ. Colorado). The surfaces of the deposits are commonly smooth, and few boulders are exposed except on steep depositional slopes or below erosional breaks in slope.

In most areas, the deposits cap high interstream divides or remnants of broad-valley erosion surfaces along the upper slopes of canyons and lie above or beyond the outer limits of younger tills (Blackwelder, 1915). Some cover extensive upland areas (Alden, 1928), but most have a patchy distribution (Capps and Leffingwell, 1904; Hole, 1910; Atwood and Mather, 1932). A zone of erratic boulders extends beyond the borders of many deposits and, in places, is all that remains of what has been inferred to have been a continuous drift sheet (Hole, 1910; Umpleby, Westgate, and Ross, 1930). This physiographic setting has suggested to many writers that the pre-Wisconsin till is precanyon in origin. The deposits tend to be dissected from 200 to over 1000 feet by modern trunk streams and to a lesser extent by an orderly tributary drainage. In some places they are so distributed that the till must have been laid down along drainage lines quite different from those existing at present (Atwood and Mather, 1932, p. 102; Atwood and Atwood, 1938a, p. 242; Richmond, 1955, U. S. Geol. Survey open file report; Ph.D. thesis, Univ. Colorado). However, some deposits have been observed on or near valley floors and locally at stream level (Holmes, 1951, p. 95; Holmes and Moss, 1955, p. 634-635; Eschman, 1955, p. 202; W. S., Cooper personal communication; J. F. Smith, Jr., and L. C. Huff, personal communication; Richmond, 1955, U. S. Geol. Survey open file report; Ph.D. thesis, Univ. Colorado).

Physical Character

The physical aspect of pre-Wisconsin till below its surficial zone of weathering is in many respects similar to that of younger tills. The deposits consist of unsorted, unsized, nonbedded, bouldery material containing soled, snubbed, faceted, or striated stones among a greater number that are angular, subangular, or subrounded. Certain differences, however, are significant. Pre-Wisconsin till in general contains a greater proportion of fine-grained constituents than younger tills in the same

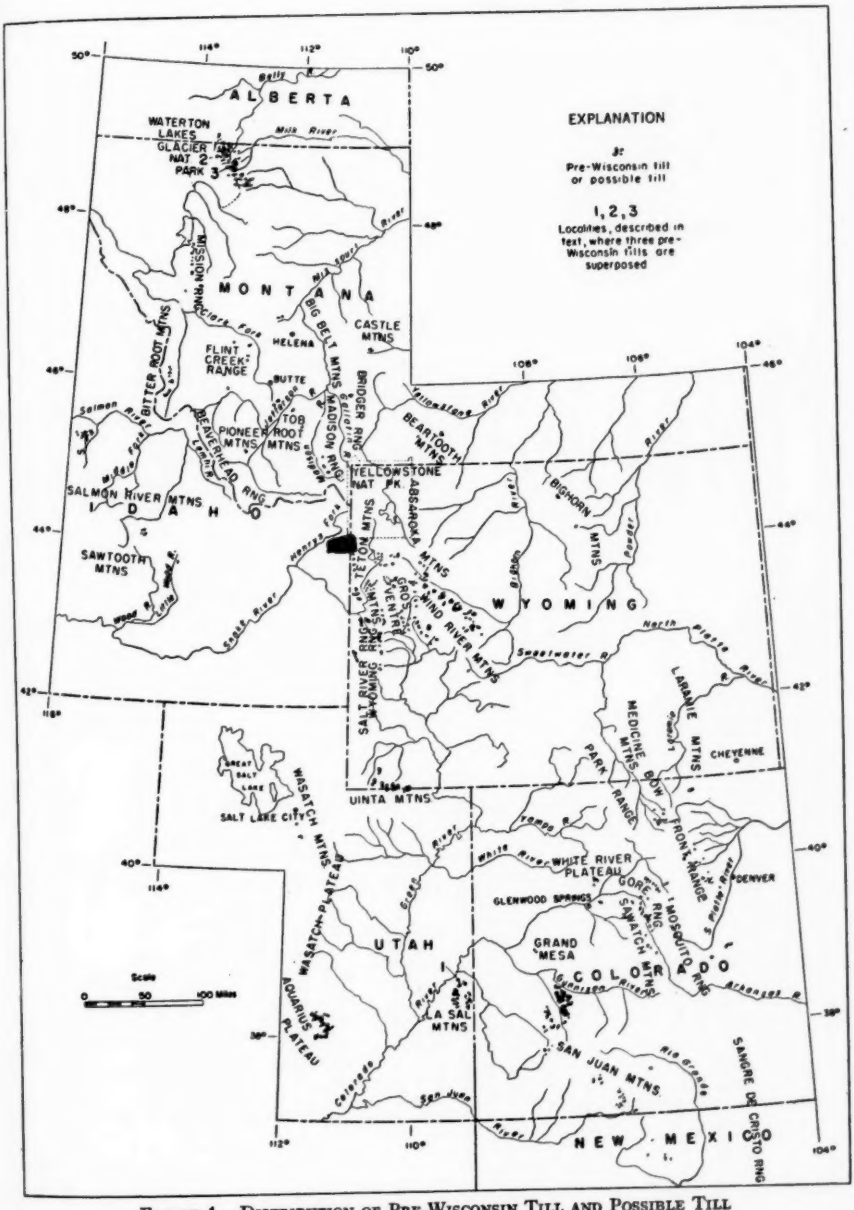

FIGURE 1.—DISTRIBUTION OF PRE-WISCONSIN TILL AND POSSIBLE TILL IN THE ROCKY MOUNTAIN REGION

TABLE 1.—DISTRIBUTION OF PRE-WISCONSIN TILL IN THE ROCKY MOUNTAIN REGION

Area	Authority	Local name	Local map	Remarks
		New Mexico		
San Juan Mountains	Ellis (1935) Smith (1936)	Cerro till	Abiquiu quad.	On Canjilon Divide
Sangre de Cristo Mountains	Ellis (1931; 1935)	Cerro till	Taos and vicinity quad.	Near towns of Red River and Taos. May not be of glacial origin (Ray, 1940, p. 1889–1891)
		Colorado		
San Juan Mountains	Hills (1884)		Montrose quad.	On Tongue Mesa
	Howe and Cross (1906, p. 260)		Ouray quad.	
	Cross, Howe, and Irving (1907, p. 7, 15)		Ouray quad.	
	Hole, (1910, p. 9)		Engineer Mountain quad.	
	Atwood and Mather (1912, p. 399–406)	San Juan drift	Montrose, Uncompahgre, Telluride, Pagosa Springs, Creede, Summitville quads.	
	Atwood and Mather (1932, p. 101–111)	Cerro till		
	Richmond (1954)		Montrose quad.	
Sawatch Mountains	Capps and Leffingwell (1904, p. 701–702)	"Older drift"	Leadville, Mt. Elbert quads.	
	Capps (1909, p. 9)	Included in "earlier drift"	Leadville, Mt. Elbert quads.	
	Powers (1935, p. 196–197)	"Earliest glacial stage, M2"	Mt. Elbert quad.	
	Nelson (1954)	Lime Creek stage	Mt. Jackson quad.	
	Tweto (1954)		Holy Cross, Minturn quads.	Deposits on northeast flank of range along drainage of Eagle River
Arkansas River Valley	Ogden Tweto (Personal communication)		Mt. Elbert, Holy Cross quads.	Deposits at several places between Tennessee Pass and Twin Lakes

Mosquito Range	Capps and Leffingwell (1904, p. 701–702)	"Older drift"	Leadville, Mt. Lincoln quads.	
	Emmons, Irving, and Loughlin (1927, p. 9)	Possible third glaciation	Leadville Mining District special map	
	Stark *et al.* (1949, p. 73–74)	Illinoian	Mt. Lincoln, Como quads.	Considered here to be of early Wisconsin age
	Singewald (1950, p. 122, 125, 129)		Mt. Lincoln, Como, Leadville quads.	
	Behre (1953, p. 16)		Mt. Lincoln, Leadville quads.	
Gore Range	Lovering and Tweto (1944, U. S. Geol. Survey open-file report)		Minturn quad.	
	Tweto (1954)		Holy Cross quad.	Deposits on west flank of range along drainage of Eagle River
	Ogden Tweto (Personal communication)		Mt. Powell, Ute Peak, Dillon quads	Deposits on east flank of range along drainage of Blue River
White River Plateau	Bass, N. W., and Richmond, G. M. (unpublished)		Glenwood Springs quad.	Upland north and south of Deep Creek
Glenwood Springs area	Bass, N. W. (Personal communication)		Glenwood Springs quad.	At Cottonwood Pass east of Glenwood Springs
Grand Mesa	Retzer, 1954	"Earliest moraine"	Grand Mesa National Forest map	On west edge of Grand Mesa along Kannah Creek
Front Range	Atwood and Atwood (1938a; 1938b)		Divide, Woodland Park quads.	On ridges near town of Divide, northwest of Pikes Peak
			Pikes Peak (1893) quad.	At a locality 10 miles southwest of Florissant
			Bailey quad.	Along U. S. Route 285, 8 miles northeast of Bailey
	Lovering and Goddard (1950, Pl. 2) Erskine, Morgan, and Robertson (1950) Boos (1954, Pl. 1) Van Tuyl, F. M. (Personal communication) Erskine, C. F. (Personal communication)		Blackhawk, Denver Mountain Parks quads.	Possible till or closely related outwash on high ridges along drainage of Clear Creek, west of Denver

TABLE 1.—*Continued*

Area	Authority	Local name	Local map	Remarks
		Colorado—*Continued*		
Front Range—*Continued*	Sheridan, D. M., and Van Horn, Richard (Personal communication)		Ralston Buttes quad.	On ridges along drainage of Ralston Creek
	Richmond, G. M. (Unpublished)		Indian Hills quad.	Outwash or possible till on flat east of Wilds Peak above Deer Creek
	Lovering and Goddard (1950, Pl. 2) Lovering and Tweto (1953, p. 26)		Tungsten quad.	Deposits of questionable origin on Tungsten Mountain
	Boos, M. F. (Personal communication)		Rocky Mountain National Park quad.	South of Estes Park
	Wahlstrom (1947, p. 562–564)	"Pre-valley" till	Tungsten, Rocky Mountain National Park quads.	Deposits on and east of Tungsten Mountain; deposits on high divides west of Ward
	Cooper, W. S. (Personal communication) Richmond (1953)		Rocky Mountain National Park quad.	Tahosa Valley, east of Longs Peak
	Bryan and Ray (1940, p. 28–29) Ray (1940, p. 1856)	Prairie Divide stage	Home, Livermore quads.	On uplands north of Cache la Poudre River
	Ives (1938, p. 1056, 1062)	Stillwater stage	Rocky Mountain National Park quad.	In Monarch Valley. Writer concurs with Ray (1940, p. 1873) that these deposits are of early Wisconsin age
	Ives (1953, p. 237, 249)	Pretty Meadow stage	Nederland quad.	In valley of North Boulder Creek. Writer believes these deposits to be of early Wisconsin age
	Lovering, T. S., and Tweto, Ogden (Personal communication)		Montezuma quad.	On west slope of range, along Snake River
Medicine Bow Mountains	Eschman (1955, p. 201–203)	Owl Mountain stage	Home quad; Jackson County Highway map	Along drainage of Michigan River; some deposits at flood-plain level

				plain level

Wyoming

Wind River Mountains	Blackwelder (1915; 1950)	Buffalo stage	Fremont Peak, Gros Ventre, Bull Lake east, Mt. Leidy quads.	
	Kirkendall (1935)	Buffalo stage	Fremont Peak quad.	
	Richmond (1941; 1948; Unpublished)	Buffalo stage	Fremont Peak, Gros Ventre, Mt. Leidy, Yountz Peak, Blue Holes, Hays Park, Burris, Bull Lake east, Bull Lake west, Wise Flat, Fort Washakie quads., Wyoming National Forest map	Two precanyon stages suggested by physiographic evidence
	Miner and Apfel (1946)	Buffalo stage	Yountz Peak quad.	Deposits on north slope of Wind River Valley, upstream from Dubois
	Moss (1951)	Buffalo stage	Wyoming National Forest map	Deposits at flood-plain level
	Holmes (1951)	Buffalo stage	Wyoming National Forest map	
	Holmes and Moss (1955)	Buffalo stage	Wyoming National Forest map	Deposits at flood-plain level
Bighorn Mountains	Salisbury and Blackwelder (1903) Blackwelder, Salisbury and Bastin (1906, p. 11) Alden (1932, p. 41)	Possible early glacial drift	Bald Mountain, Dayton, Cloud Peak, Fort McKinney quads.	Many of deposits are early Tertiary fanglomerate (Sharp, 1948)
Yellowstone National Park	Alden (1928)	Buffalo stage	Shoshone, Lake quads.	Very extensive in southwestern part of Park; glacial origin of deposits along Yellowstone River questioned by Howard (1937)
Teton Mountains	Blackwelder (1915) Fryxell (1930) Horberg (1938) Edmund (1951)	Buffalo stage	Grand Teton, Mt. Leidy quads.	
Gros Ventre Range	Horberg (1938) Nelson and Church (1943)	Buffalo stage	Gros Ventre quad.	

TABLE 1.—*Continued*

Area	Authority	Local name	Local map	Remarks
		Wyoming—*Continued*		
Wyoming and Salt River Ranges	Alden, W. C. (Personal communication) Rubey, W. W. (Personal communication)		Afton, Cokeville quads.	
Uinta Mountains	Bradley (1936) Richmond, G. M. (Unpublished)	Little Dry stage	Topographic map (Bradley, 1936)	
Medicine Bow Mountains	Atwood (1937)		Medicine Bow quad.	Believed to be of early Wisconsin age by Ray (1940) and Mears (1953)
		Montana		
Glacier National Park	Alden (1912; 1924; 1932) Alden and Stebinger (1913) Horberg (1954, p. 1102) Richmond, G. M. (Unpublished)	Illinoian or Iowan (?) stage Kennedy drift	Little Chief, Browning quads. Browning quad. Little Chief, Browning quads.	On high divides on east slope of park. Two possible stages Three tills superposed
Waterton Lakes area, Alberta	Alden (1932, p. 32–33) Horberg (1954, p. 1102)	Kennedy drift	Lethbridge, sectional map no. 15, Alberta	
Beartooth Mountains	Bevan (1927) Alden (1932, p. 41)		Mt. Wood quad.	
Castle Mountains	Weed (1896, p. 144–145; 1899, p. 3, 7) Alden (1932, p. 40; 1953, p. 62)		Jeffersonal National Forest map	South flank of mountains west of Lennep
Big Belt Mountains	Pardee (1925, p. 37) Alden (1953, p. 62)		Duck Creek Pass quad.	
Bridger Range; Madison Range; Tobacco Root Mountains	Alden (1953, p. 63–64)		Belgrade quad; Madison National Forest map; Gallatin National Forest map	

Mountains				
Pioneer Mountains	Pardee (1951, p. 82) Alden (1953, p. 64)		Beaverhead National Forest map	
Butte region	Atwood and Atwood (1938a; 1938b) Alden (1953, p. 64)		Butte special topographic map	Vicinity of Grace
Helena region	Alden (1953, p. 62)		Helena, Marysville quads.	North rim of Little Prickly Pear Creek
Flint Creek range	Alden (1953, p. 66)		Drummond, Phillipsburg quads.	
Bitteroot Mountains	Alden (1953, p. 67) Ross, C. P. (Personal communication)		Hamilton quad.	On east flank of mountains
Mission Range	Alden (1953, p. 67) Stebinger, Eugene (Personal communication)		Flathead National Forest map	Along canyon rim on west slope of mountains
		Idaho		
West Slope Teton Mountains	Blackwelder (1915, p. 330)	Buffalo stage	Targhee National Forest and Teton County maps	On divides between North Fork of Teton River and Warm River
Salmon River	Ross (1929; 1937)		Bayhorse quad.	
Middle Fork, Salmon River	Ross (1934)		Casto quad.	
South Fork, Salmon River	Capps (1940; 1941)		Idaho National Forest map	
Little Wood River	Umpleby, Westgate, and Ross (1930)		Hailey quad.	
Henrys Fork, Snake River	Stearns, Bryan, and Crandall (1939)		Targhee National Forest and Teton County maps	Extension of deposits on western slope of Yellowstone National Park and Teton Mountains, Wyoming

TABLE 1.—*Concluded*

Area	Authority	Local name	Local map	Remarks
		Utah		
Wasatch Mountains	Atwood (1909)	Earlier glacial epoch	Fort Douglas quad.	Considered of early Wisconsin age by Blackwelder (1931)
	Ives (1950)	Graniteville erratics	Draper quad.	At mouth of Little Cottonwood Canyon
	Crittenden, M. D., and Marsell, R. E. (Personal communication)		Draper quad.	Erratics at mouth of Little Cottonwood Canyon
	Hunt, Varnes, and Thomas (1953)		Timpanagos Cave quad.	Near mouth of Dry Creek northeast of Alpine
Wasatch Plateau	Spieker, E. M. (Personal communication)		Manti National Forest map	Possible till
Aquarius Plateau	Smith, J. F., and Huff, L. C. (Personal communication)		Loa 1, 4; Notom 2, 3 quads.	Extensive deposits of till, possible till and associated landslide debris
La Sal Mountains	Richmond (1952) Richmond (1955, Open file report, U. S. Geol. Survey; Ph.D. thesis, Univ. Colo.)	"Two earlier stages" Harpole Mesa formation	Mt. Peale 1, 2; Mt. Wass 3, 4 quads.	Three tills superposed

area. It also tends to be more compact than younger tills and in some places has a well-developed blocky jointing along which a dark stain, possibly manganese bearing, is developed. Jointing and staining are uncommon in younger tills.

In some areas, pre-Wisconsin till is reported to differ lithologically from younger tills. For example, in the San Juan Mountains, the Cerro till locally contains rock types believed to have been completely stripped from the source area before deposition of the younger Durango till (Atwood and Mather, 1932, p. 107; Atwood and Atwood, 1938a, p. 240); in the Sawatch Range near the head of the Eagle River, pre-Wisconsin till lacks rock types present in younger tills, probably because these types were not exposed until the intervening erosional interval (Ogden Tweto, personal communication).

Disintegrated, rotted, or crumbly cobbles or boulders are reported by many writers to be abundant in the till, especially in deposits containing crystalline rocks. However, in areas such as the Uinta Mountains where most of the till is composed of resistant quartzite, very little rotted material is noted (Bradley, 1936). Decomposed rock is commonly much more abundant in the weathered zone at the top of the till than at depth, and, in some places certain rock types have been entirely dissolved from the weathered zone. However, in some places decomposed rock is abundant throughout the deposits. Some of the decomposition at depth appears to be due to subsoil decay of the stones *in situ*, even where fresh and rotted stones of the same rock type occur in close proximity. However, in areas such as the Wind River Mountains of Wyoming and the Front Range of Colorado, many exposures of deeply decomposed bedrock beneath the till suggest that much of the rotted material in the till has been derived from this source and is of preglacial origin.

Surficial Weathering and Soil Development

Although boulders at the surface of the till are abundant only locally, they are considerably more weathered than those on younger tills. The more resistant rock types are generally extensively fractured and in places are deeply sand-blasted, grooved, or polished. Crystalline rocks in relatively humid environments tend either to be decomposed throughout or to have hard, iron-stained cores beneath decomposed or exfoliated exteriors. In relatively arid environments, some apparently sound boulders have a disintegrated interior beneath a hard exterior shell.

Resistant rocks such as quartz, quartzite, or siliceous argillite rocks tend to be concentrated at the surface in greater abundance than at depth, and readily soluble rock such as limestone, dolomite, calcareously cemented sandstone, or granite (Capps, 1940, p. 13–14) are lacking at the surface in some places. For example, in the La Sal Mountains of Utah, quartz, which is rare at the surface of Wisconsin deposits, is almost 10 times as abundant at the surface of pre-Wisconsin deposits; and east of Glacier National Park in Montana, certain pre-Wisconsin gravel deposits that contain considerable amounts of limestone, dolomite, and diorite at depth, consist almost wholly of siliceous argillite at surface exposures. Along the northwest slope of the Wind River Mountains in Wyoming, quartzite cobbles derived from a conglomerate of Paleocene age are more abundant at the surface of pre-Wisconsin deposits than at depth.

Most writers have reported that pre-Wisconsin till has a thick zone of intense weathering at its surface. This zone is an ancient soil profile or paleosol (Hunt and Sokoloff, 1950) that formed on the till during one or more interglacial intervals of pre-Wisconsin time. In many places this soil profile is buried beneath Wisconsin and younger deposits; elsewhere it is relict at the surface and truncated by younger erosion surfaces on which Wisconsin and younger deposits rest.

Like stratigraphically younger soils, the pre-Wisconsin soil changes markedly with environment, and its broader characteristics are geographically related to climatic provinces. For example, in the La Sal Mountains, Utah, the writer (Richmond, 1955, U. S. Geol. Survey open-file report; Ph.D. thesis, Univ. Colorado) traced the soil on pre-Wisconsin till and associated outwash from a Brown soil facies at lower altitudes through a Brown Forest facies at intermediate altitudes to a Brown Podzolic facies

in the higher parts of the mountains; the writer has also observed similar facies changes in pre-Wisconsin soils in the region of Glacier National Park, Montana. Pre-Wisconsin Brown soil profiles have been described from the base of the Wasatch Mountains, Utah (Hunt and Sokoloff, 1950; Hunt, Varnes, and Thomas, 1953), from the east base of the Front Range, Colorado (Hunt and Sokoloff, 1950; Hunt, 1954), and at other isolated localities. The profile is characterized by a reddish leached horizon of clay accumulation (*B horizon*) that commonly displays a well-developed soil structure and is 4–10 feet thick. This overlies a whitish strongly impregnated to well-cemented horizon of calcium carbonate accumulation (*Cca horizon*) that ranges from about 8 to 15 feet in thickness and is the "caliche zone" of many writers.

Brown Forest profiles have been described only from the La Sal Mountains (Richmond, 1955, U. S. Geol. Survey open file report; Ph.D. thesis, Univ. Colorado) but have been observed by the writer in the Front and Sawatch Ranges in Colorado, the Wind River and Teton Mountains in Wyoming, and the mountains of Glacier National Park in Montana. The profile is similar to that of the Brown soil facies, except that the horizon of calcium carbonate accumulation is much less impregnated with calcium carbonate and is rarely cemented.

Profiles considered by the writer to be Brown Podzolic profiles have been noted in the Wind River Mountains in Wyoming, the Front Range and San Juan Mountains in Colorado (Richmond, 1953; 1954), and the La Sal Mountains in Utah (Richmond, 1955, U. S. Geol. Survey open file report; Ph.D. thesis, Univ. Colorado). They have also been observed by the writer in the Wasatch Mountains, the Sawatch Mountains, the Uinta Mountains, the Teton Mountains, and Glacier National Park. Probably most of the deeply weathered and oxidized zones on pre-Wisconsin till throughout the higher parts of the Rocky Mountains, noted by many writers, are Brown Podzolic soils. These profiles are characterized by a thick reddish to reddish-brown horizon of clay accumulation (*B horizon*) that ranges in thickness from 4 to about 15 feet and is gradational downward into the parent till. A well-developed soil structure is displayed except in very bouldery deposits, and a large part of the less-resistant rock constituents of the till are deeply decomposed or leached.

In general, the profile characteristics of pre-Wisconsin soils are much more strongly developed than those of stratigraphically younger soils: the *B horizons* are thicker, redder, more clayey, and have a more definite soil structure; the *Cca horizons* are thicker and more strongly impregnated with calcium carbonate. Decomposition of parent material throughout the profile is also markedly greater. However, the pre-Wisconsin soil is crosscut by younger erosion surfaces so that in many places it is partly or wholly stripped from the underlying till, and a younger less well-developed soil has formed in its stead. Complete profiles are difficult to find, and careful attention to evidence of their stratigraphic position as well as to their profile characteristics is necessary to identify them.

Evidence for pre-Wisconsin Age of the Till

Almost all who have observed the characteristics of the till have considered it to be of pre-Wisconsin age; this conclusion rests on a variety of evidence, which though reasonable is not incontestable. The deposits are clearly older stratigraphically than till believed by various workers to be of either Illinoian or early Wisconsin age, because in many places they are overlain by the younger till and separated from it by a zone of weathering or are truncated by an erosion surface on which the younger till rests. However, the evidence for the age of the younger till is not conclusive.

The topographic position of many deposits on divides above and beyond the outer limits of younger tills, their degree of erosion and locally inconsistent relations to present drainage, and the degree to which morainal forms have been destroyed suggests considerable age. The presence of some deposits in valley bottoms and of others with distinct morainal form, however, suggests that their age, at least locally, may not be great. Some writers have inferred that such deposits could be of Wisconsin age (Holmes, 1951).

No material suitable for radiocarbon dating

has been found in the till, and the age, if pre-Wisconsin, would exceed the range of such dating by present methods. Few fossils have been found in the till or in associated outwash gravel. Mammoth remains of possible Yarmouth age are reported from the Cerro till in Colorado (Cook, 1930; 1931); bones of musk ox and giant bison, considered to be of Yarmouth age (Hay, 1924, p. 182, 195), and bones of horse of possible Yarmouth age (Hunt, 1954) have been found in possible outwash deposits near Denver and Colorado Springs. Mammoth remains of Yarmouth age are also known from possible outwash deposits in Idaho (Hay, 1927, p. 43, 44, 126, 127).

Next to geomorphic setting, the criteria that are most widely applied as evidence that the till is probably of pre-Wisconsin age are the depth and degree of weathering that have affected it. This kind of weathering as described generally in the literature (Hunt and Sokoloff, 1950; Hunt, Varnes, and Thomas, 1953; Richmond, 1955, U.S. Geol. Survey open-file report; Ph.D. thesis, Univ. Colorado) has produced much more strongly developed soil profiles than those found on younger tills. The profiles are, moreover, remarkably similar to those developed on pre-Wisconsin till in the mid-continent region. For example, the horizon of clay accumulation (*B horizon*) in the Brown soil facies is so similar in its reddish color, amount of clay, soil structure, and over-all thickness to that of well-drained soils on pre-Wisconsin till in Kansas and Nebraska, that except for lithologic differences in the contained rock fragments samples from each are indistinguishable. The color, consistence, soil structure, thickness, and degree of calcium carbonate impregnation and cementation in the horizon of calcium carbonate accumulation are also similar. Profiles of the Brown Podzolic facies on the till in the mountains are also remarkably similar to well-drained Brown or Grey Brown Podzolic profiles developed on pre-Wisconsin till in Iowa and Illinois.

Physiographic Evidence Suggesting More Than One pre-Wisconsin Glaciation

Blackwelder (1915, p. 330) pointed out that more than one episode of pre-Wisconsin glaciation might be encompassed by deposits included in the Buffalo glacial stage of Wyoming. East of the Wind River Mountains, along the Wind River southeast of Bull Lake, the writer observed that gravel deposits on two high terraces grade headward into Buffalo till (Richmond, 1948, p. 1401). A thick and very strongly developed Brown soil is formed on both the gravel and the till. Although no critical evidence that interglacial conditions separated two episodes of glaciation was noted, both terrace gravels are precanyon in age and are separated by an erosion cycle of lesser magnitude. West of the Wind River Mountains, along the drainage of the Green River, gravel deposits on a terrace 130–195 feet above stream level are outwash from Buffalo till; gravel deposits on a terrace 100–135 feet above stream level are not in contact with Buffalo till, and Buffalo till occurs at stream level (Holmes and Moss, 1955, p. 634). In an adjacent area to the north, Buffalo till is extensively preserved on high interfluves (Blackwelder, 1915; Kirkendall, 1935; Richmond, 1948) and in the same area has been observed by the writer only a few feet above the Green River near Daniel and Bronx. These facts suggest one postcanyon and two precanyon glaciations but could also be explained as recessional phases of a single stage or by uplift of the mountains during glaciation, as pointed out by Holmes and Moss (1955, p. 638).

Alden and Stebinger (1913) discussed the relations of gravel deposits on two high erosion surfaces in Montana, east of Glacier National Park, to deposits of pre-Wisconsin till and concluded that two episodes of glaciation might be indicated. Later, Alden (1924, p. 405; 1932, p. 31–32) found striated cobbles in the gravels on both surfaces and noted that the higher or No. 1 bench was capped by till near the mountains; he again postulated two pre-Wisconsin glacial stages but cautioned that the conditions could be accounted for by a single drift. The writer recently found a third pre-Wisconsin erosion surface which lies below the No. 2 bench of Alden (1932) and is capped by gravel that grades into till near the mountains.

In the La Sal Mountains, Utah, thick deposits of pre-Wisconsin till and associated outwash gravel locally cap each of two extensive high erosion surfaces at the mountain front

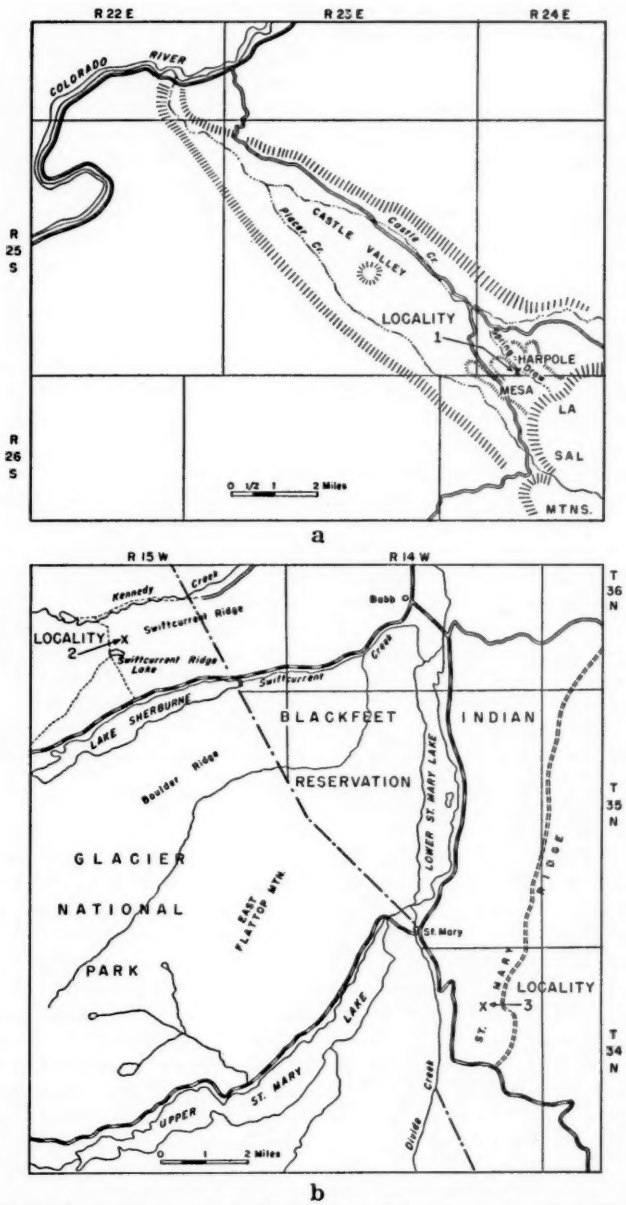

FIGURE 2.—MAPS SHOWING LOCATIONS OF EXPOSURES OF THREE SUPERPOSED PRE-WISCONSIN TILLS
A. Harpole Mesa, La Sal Mountains, Utah. B. Near Glacier National Park, Montana

(Richmond, 1955, U. S. Geol. Survey open-file report; Ph.D. thesis, Univ. Colorado). The lower surface has been cut more than 200 feet below deposits of till on the upper surface and is clearly younger than that till. A third, younger pre-Wisconsin till forms large but poorly preserved moraines on remnants of a broad-valley surface preserved along the upper slopes of canyons that dissect the lower of the two high surfaces. This physiographic evidence strongly suggests three pre-Wisconsin stages of glaciation.

Stratigraphic Evidence for Three Pre-Wisconsin Glacial Stages

General Statement

Evidence for three pre-Wisconsin glaciations separated by interglacial intervals of weathering and soil formation has been found by the writer in three stratigraphic sections. One section is in the La Sal Mountains of southeastern Utah; the other two are near the eastern boundary of Glacier National Park in Montana (Fig. 2).

La Sal Mountains, Utah

At the northwest edge of the La Sal Mountains (Fig. 2a, locality 1), a graben of Tertiary origin underlies the present site of Castle Valley. Renewed subsidence of this trough in Pleistocene time locally reduced the effects of erosion so that deeply weathered till and associated deposits of a first glacial stage were overlapped by till and outwash of a second stage. After deep weathering, these deposits were dissected as much as 200 feet. Later, they were partly overlapped by till and outwash of a third pre-Wisconsin stage which, in turn, were deeply weathered before final canyon erosion and subsequent Wisconsin glaciations.

A good section of these deposits is exposed on the south slope of Spring Draw on Harpole Mesa at the mountain front (SW¼ sec. 32, T. 25 S., R. 24 E.). Here, the oldest pre-Wisconsin till, with associated eolian and alluvial deposits, rests unconformably on a folded conglomerate of Pliocene (?) or very early Pleistocene (?) age and has a deeply weathered Brown soil profile at its top. Above this soil lies a second till, on which a second deeply weathered Brown soil is developed. At the top of the section is a solifluction deposit on which a third deeply weathered Brown soil profile is formed. The solifluction deposit, ranging from 3 to 15 feet in thickness, can be traced eastward along the edge of Harpole Mesa into a saddle that dissects the middle till. Here it grades into a third pre-Wisconsin till and associated outwash deposits on which a deeply weathered Brown soil profile is also formed. On the south rim of Harpole Mesa, other less well-exposed sections of this succession of deposits overlook the canyon of Placer Creek, on the floor of which are moraines and outwash deposits of Wisconsin age.

Section 1.[1] *South side of Spring Draw on Harpole Mesa, La Sal Mountains, Utah (SW ¼ sec. 32, T. 25 S., R. 24 E.)*

	Thickness (Feet)
Pre-Wisconsin soil No. 3 on solifluction mantle	
B horizon: reddish-brown (2.5 YR-5/4), unsorted, nonbedded, arkosic, sandy loam with angular to subangular rock fragments, dominantly diorite and sandstone. Many fragments broken since deposition; some deeply disintegrated. *Structure:* moderate, medium columnar. *Consistence:* slightly sticky, slightly plastic, friable, slightly hard. *Reaction:* pH 7.0	3
Abrupt irregular boundary	
Cca horizon: pink (5 YR-7/3), unsorted, arkosic, sandy solifluction mantle containing angular to subangular rock fragments, dominantly diorite and sandstone. Many fragments broken, some deeply disintegrated. *Structure:* structureless to weak, thick platy. *Consistence:* nonsticky, nonplastic, firm, slightly hard. *Reaction:* pH 8.5	3
Total thickness of solifluction mantle	6
Gradual contact	
Pre-Wisconsin soil No. 2 on till No. 2	
B horizon: light-reddish-brown (2.5 YR-6/4), unsorted, arkosic, sandy till, containing subangular to subround fragments of diorite and sandstone. *Structure:* moderate, very coarse angular blocky. *Consistence:* slightly sticky, slightly plastic, firm, hard; weakly cemented with clay and iron oxide. *Reaction:* pH 7.5. Interior of joint	

[1] Soil profile descriptive terminology of this and subsequent sections follows usage of U. S. Dept. Agriculture, Soil Survey Manual (1951).

	Thickness (Feet)
blocks leached of carbonate; secondary carbonate, probably derived from soil No. 3 along structure planes	4
Abrupt contact	
Cca horizon: pink (7.5 YR-7/4), unsorted, arkosic, sandy till, containing subangular to subground fragments of diorite and sandstone. *Structure:* strong, very coarse platy. *Consistence:* nonsticky, nonplastic, firm, hard; strongly cemented with calcium carbonate. *Reaction:* pH 8.5	15
Diffuse boundary	
C horizon: light-yellowish-brown (10 YR-6/4) till consisting of unsorted, nonbedded, arkosic sandy silt and subangular to round fragments of diorite and sandstone. Boulders up to 6 feet in diameter. Compact, but friable. *Reaction:* pH 7.5. Some pods and veins of powdery calcium carbonate up to 2 inches in diameter	28
Total thickness of till No. 2	47
Erosional contact	
Pre-Wisconsin soil No. 1 on till No. 1	
B horizon: light-reddish-brown (5 YR-6/4), nonbedded, well-sorted eolian fine sand and silt. *Structure:* moderate, very coarse angular blocky. *Consistence:* slightly sticky, slightly plastic, firm, hard. *Reaction:* pH 7.5 except for concentrations of secondary carbonate along structure planes	6
Abrupt boundary	
Cca horizon: light-brown (7.5 YR-6/4), unsorted, arkosic, sandy till containing abundant subangular to round fragments of diorite and sandstone. *Structure:* moderate, very coarse platy. *Consistence:* nonsticky, nonplastic, firm, hard. *Reaction:* pH 8.5; moderately cemented with calcium carbonate	7
Abrupt contact	
Cca horizon, continued: pink (7.5 YR-7/4), well-sorted, nonbedded eolian sand and silt. *Structure:* very coarse platy. *Consistence:* nonsticky, nonplastic, firm, hard. *Reaction:* pH 8.5	2
Abrupt contact	
Till: light brown (7.5 YR-6/4), sandy; contains abundant subangular to round fragments of diorite and sandstone. *Reaction:* pH 7.5	8
Abrupt contact	
Alluvial sand and silt: light-reddish-brown (5 YR-6/4), crudely bedded; contains local bedded-gravel lenses. *Reaction:* pH 7.5	1.5
Abrupt contact	
Till: light-yellowish-brown (10 YR-6/4), unsorted, nonbedded, arkosic sand and silt, containing subangular to round fragments of diorite and sandstone. Boulders up to 5 feet in diameter. Structureless, compact, and friable. *Reaction:* pH 7.5. A few thin lenticular layers of platy carbonate in upper few feet	29
Abrupt contact	
Eolian sand and silt: light-reddish-brown (5 YR-6/4), massive, nonbedded. Contains a few small stones. *Reaction:* pH 7.0 to 7.5	2
Total thickness of lower till and associated deposits	55
Total thickness of pre-Wisconsin deposits at this locality	108
Angular unconformity	
Conglomerate of Pliocene (?) or very early Pleistocene (?) age	

Glacier National Park, Montana

Two exposures on the east slope of the mountains in Glacier National Park, Montana, display evidence of multiple pre-Wisconsin glaciation. One is on Swiftcurrent Ridge, the other on St. Mary Ridge (Fig. 2b, localities 2, 3).

The exposure on Swiftcurrent Ridge is in a landslide scarp overlooking the canyon of Kennedy Creek northeast of Swiftcurrent Ridge Lake (SE¼, sec. 33, T. 36 N., R. 15 W.) The exposure is about 75 feet high and discloses three superposed pre-Wisconsin tills, on each of which is a thick very deeply weathered Brown Podzolic soil profile. The base of the lower till is not exposed.

Section 2. North rim of Swiftcurrent Ridge near east boundary of Glacier National Park, Montana (SE ¼ sec. 33, T. 36 N., R. 15 W.)

	Thickness (Feet)
Wisconsin and Recent soil complex	
A horizon: dark-gray organic and stony mineral matter	0.3
B horizon: brown (10 YR-5/3), stony, sandy frost mantle derived from underlying till. Angular to subangular rock fragments of limestone, red and green argillite, quartzite, and diorite. Some soled and faceted, some striated. Includes rotted fragments of all lithologic types, but most abundant are diorite and sandy limestone. *Structure:* moderate, medium subangular blocky. *Consistence:* sticky, plastic, friable, slightly hard. *Reaction:* pH 5.0	2.5
Abrupt, irregular, and locally involuted boundary	

	Thickness (Feet)	
Pre-Wisconsin soil No. 3 on till No. 3		
B horizon: reddish-brown (5 YR-4/3 to 4/4) clayey bouldery till, lithologically like that of *C horizon* beneath but containing much more clay and more rotted fragments. *Structure:* strong, medium subangular blocky. *Consistence:* sticky, plastic, very firm, hard. *Reaction:* pH 5.0.....	6.0	
Diffuse boundary		
C horizon: yellowish-red (5 YR-4/8), sandy, bouldery till containing angular to subangular fragments of limestone, red and green argillite, quartzite, and diorite. Some fragments soled and faceted, some striated; many rotted, especially those of diorite and sandy limestone. Much more bouldery than tills 1 and 2. *Structureless. Consistence:* nonsticky, nonplastic, friable, slightly hard. *Reaction:* pH 5.0.....	9	
Total thickness of till No. 3.....		15
Abrupt irregular erosional contact with material from beneath locally mixed with till above		
Pre-Wisconsin soil No. 2 on till No. 2		
B_2 horizon (upper part): yellowish-red (5 YR-4/8–5/8), stony, sandy, clay till, lithologically like *C horizon* beneath but containing more clay and more rotted fragments. More clayey and redder than *B_3 horizon. Structure:* strong, coarse subangular blocky. *Consistence:* very sticky, very plastic, firm, hard. *Reaction:* pH 5.0 (top); 5.5 (bottom).....	8	
Gradational boundary		
B_3 horizon: strong-brown (7.5 YR-5/6), stony, sandy, clay till, lithologically like *C horizon* beneath but more clayey. *Structure:* moderate, fine subangular blocky. *Consistence:* slightly sticky, plastic, firm, hard. *Reaction:* pH 5.5.....	4	
Diffuse boundary		
C horizon: red (2.5 YR-4/6), sandy, stony till containing gravel- to boulder-sized fragments of limestone, red and green argillite, quartzite, and diorite. Some fragments soled and faceted; some striated; a few rotted stones, especially those of diorite and sandy limestone. Relatively fewer cobbles and boulders than in tills 1 or 3. Local massive layers of yellowish-red (7.5 YR-6/6), clean, fine sand, 4–12 inches thick, in lower part. *Structureless. Consistence:* nonsticky, nonplastic, friable, slightly hard. *Reaction:* pH 6.0....	14	
Total thickness of till No. 2.....		26
Abrupt irregular erosional contact with material from till beneath locally mixed with till above		
Pre-Wisconsin soil No. 1 on till No. 1		
Altered B_2 horizon: red (10 R-4/8), stony, sandy clay till, mottled strong brown (7.5 YR-5/8). Mottles become fainter, smaller, and more widely spaced downward. Lithologically similar to that of *C horizon* below but more clayey and contains more rotted rock fragments. *Structure:* strong, medium subangular blocky. *Consistence:* sticky, plastic, firm, hard. *Reaction:* pH 5.5 (red), pH 5.0 (yellow)....	7	
Diffuse boundary		
B_2 horizon: red (10 R-4/6), stony, sandy clay till, lithologically similar to that of *C horizon* beneath, but more clayey and contains more rotted rock fragments. *Structure:* moderate, medium subangular blocky. *Consistence:* sticky, plastic, firm, hard. *Reaction:* pH 5.5.....	7	
Diffuse boundary		
C horizon: weak-red (10 R-5/4), stony, sandy, silt till containing angular to subangular fragments of limestone, red and green argillite, quartzite, and diorite. Some fragments soled and faceted, some striated; some rotted, especially those of diorite and sandy limestone. Most fragments gravel-sized, but cobbles and boulders numerous. Color of till due to large amount of finely divided red argillite in matrix. *Structure:* weak, medium subangular blocky in upper part, structureless in lower part. *Consistence:* slightly sticky, slightly plastic, friable, hard. *Reaction:* pH 6.0.....	15 exposed	
Total exposed thickness of till No. 3.		29
Total thickness of section.....	70	

The exposure on St. Mary Ridge is at the head of a large landslide scarp on the northwest slope of the ridge (center, sec. 11, T. 34 N., R. 14 W.), several miles southeast of the exposure on Swiftcurrent Ridge in an environment in which Brown soils occur. The deposits were discussed by Alden (1932, p. 35–36), who recognized 30–50 feet of Wisconsin gray till capping an oxidized weathered zone on about 200 feet of pre-Wisconsin till that is in part well cemented with calcium carbonate. They have also been examined by Horberg (1954, p. 1125). Detailed study of these deposits by the

writer and R. B. Morrison indicates that the Wisconsin till is of early Wisconsin age, and that the lower deposits include at least three pre-Wisconsin tills, each with a deeply weathered Brown soil profile.

Section 3. Head of large landslide scar on northwest slope of St. Mary Ridge (center, sec. 11, T. 34 N., R. 14 W.)

	Thickness (Feet)
Early Wisconsin till:	
Pinkish-gray (7.5 YR-7/3 to 7/4), loose to slightly compact, stony, sandy till. Unsorted, unsized, structureless. Stones up to 10 feet in diameter, but most less than 6 inches, rounded to angular, most subangular. Mountain rock types including limestone. Some diorite rotted, some fresh. Deposit contains many chunks of lime-cemented till derived from underlying pre-Wisconsin till. Fine-grained component is nonsticky, nonplastic, sandier than pre-Wisconsin till. Lower 15 feet more brownish than upper part. *Reaction:* pH about 8.0. Deposit is thickest in western part of exposure.....	30–50
Abrupt erosional disconformity	
Pre-Wisconsin soil No. 3 on till No. 3	
B horizon: Light reddish-brown to reddish-yellow (7.5 to 5 YR-6/4 to 6/6), stony, clayey till. *Structure:* strong, medium, angular, blocky. *Consistence:* sticky, plastic, hard, compact. Till contains estimated 20 per cent stones greater than 2 inches in diameter, 50 per cent stones less than 2 inches in diameter, 30 per cent sand, silt, and clay. Composed of mountain rock types, but all limestone leached. Much of diorite rotted. *Reaction:* pH 6.0. Horizon cut out by erosion in eastern part of exposure.....	6–12
Gradual boundary	
Cca horizon: Pink (7.5 YR-7/3), stony till as below, firmly cemented with calcium carbonate. Cementation tends to be in layers that strike N. 20°–30° W. and dip 25°–30° E. and are unrelated to structure of till. Some calcium carbonate precipitated from water moving laterally through deposit. *Reaction:* pH 8.5.....	15–20
Diffuse boundary	
Till No. 3: Pink (7.5 YR-7/3), compact, stony till. Unsorted, unsized, structureless. Estimated composition, 20 per cent stones greater than 2 inches in diameter, 30 per cent cent stones less than 2 inches in diameter, 50 per cent sand and finer material. Most stones subangular, some soled, snubbed, faceted, or striated. Lithologically all mountain rocks, including limestone and diorite. Some of latter rotted. Matrix compact, firm, slightly sticky, slightly plastic. *Reaction:* pH 8.0.....	10
Total thickness of till No. 3....	31–42
Abrupt erosional disconformity	
Pre-Wisconsin soil No. 2 on till No. 2	
B horizon: reddish-brown (5 YR to 7.5 YR-5/4 to 5/6), stony, clayey till. *Structure:* strong, medium angular blocky. *Consistence:* compact, sticky, plastic, firm. Uncemented till as below, except leached of carbonate and containing no limestone, though soft, sandy, noncalcareous relicts of former limestone present. Much of diorite rotted. *Reaction:* pH 6.5 at top, 7.0 at bottom.....	13
Gradual boundary	
Cca horizon: Pink (7.5 YR-7/3), stony, sandy till as below, but firmly cemented with calcium carbonate. Cementation tends to have layering that strikes N. 20°–30° W. and dips 25°–30° E., unrelated to structure of till. No drusy coating noted, but much of carbonate probably precipitated from water moving laterally through deposit. *Reaction:* pH 8.0.....	20–25
Abrupt boundary	
Sand: yellow (10 YR-7/6), well sorted, medium-grained, loose. *Reaction:* pH 8.0.....	3
Till No. 2: pink (7.5 YR-7/3), stony, sandy till. Estimated textural composition: 20 per cent stones greater than 2 inches in diameter, 30 per cent stones less than 2 inches in diameter, 50 per cent sand, silt, and clay. Compact, firm, sticky, plastic. Some stones snubbed, soled, faceted, or striated. All lithologically mountain types, including limestone. Some, of diorite, rotted. *Reaction:* pH 8.0.....	15
Gradual boundary	

	Thickness (Feet)	
Partially leached zone: yellow (10 YR-7/6), stony, sandy till, like above, but less clayey and partly leached of carbonate. Loose to slightly compact, slightly sticky, slightly plastic. *Reaction:* pH 6.5	7	
Gradual boundary		
Till No. 2: pink (7.5 YR-7/3), stony, sandy till. Uncemented, but locally containing angular chunks of cemented till. *Reaction:* pH 8.0	16	
Total thickness of till No. 2		74–79
Erosional contact		
Pre-Wisconsin soil No. 1 on till No. 1		
B horizon: yellowish-red (5 YR-5/6), stony, clay till. *Structure:* strong, medium angular blocky. *Consistence:* sticky, plastic, compact, hard. Leached of carbonate. No limestone, except rotted noncalcareous sandy relicts. Most diorite rotted. *Reaction:* pH 6.5	0–5	
Abrupt boundary		
Cca horizon: pink (7.5 YR-7/3), hard till, strongly cemented with calcium carbonate in layers that strike N. 20°–30° W. and dip 25°–30° E. Some interstices, incompletely filled with carbonate, have fine drusy lining. Much of carbonate probably precipitated from water moving laterally through deposit. Till unsorted, unsized, and nonbedded. Contains a few thin, bedded, silty lenses. Estimated textural composition: 30 per cent sand, silt, and clay, 50 per cent stones 2 inches in diameter and smaller, 20 per cent stones larger than 2 inches in diameter. Largest boulders about 5 feet in diameter. Many stones soled, snubbed, faceted, or striated. Stones rounded to angular, most subangular. Rock types include red and green argillite (most abundant), sandstone, quartzite, limestone, and diorite; all from mountains. Some of diorite rotted. *Reaction:* pH 8.0	40	
Abrupt contact		
Till No. 1: Most covered by talus and wash debris. Deposit poorly exposed but lithologically and texturally like *Cca horizon* above. Compact and calcareous but not strongly cemented with calcium carbonate	35	
Total thickness of till No. 1		75–80
Total thickness of composite section		210–251

Figure 3 is a sketch of the exposure, showing the continuity of the units described above. Three pre-Wisconsin tills, separated by zones of weathering, are apparent. The significance of the "partly leached" zone in the lower part of till No. 2 is not clear. It could represent another zone of weathering but it could also represent local subsurface leaching, possibly related to water moving through the overlying sand. Its characteristics are not like those of the *B-horizons* of other weathered zones in the section. For example, it is yellow rather than reddish and is more loose and less plastic than those horizons. Lacking continuity and more distinctive characteristics, it cannot at present be considered stratigraphically significant.

Correlation

Many have hazarded tentative statements regarding the possible correlation of pre-Wisconsin till in the Rocky Mountain region with one or more of the three pre-Wisconsin stages in the Mississippi Valley. In Colorado, the Cerro till has been considered Kansan or possibly Nebraskan (Atwood and Mather, 1932; Atwood and Atwood, 1938). In Wyoming, the Buffalo stage has been compared to the Kansan or possibly the Illinoian stage (Blackwelder, 1915; 1931); Kansan is preferred by most writers (Fryxell, 1930; Bradley, 1936; Horberg, 1938). In Montana, the deposits have been regarded as representing two possible stages, the Nebraskan and the Kansan (Alden and Stebinger, 1913, p. 570; Alden, 1924, p. 405–406; 1932, p. 31–32; 1953)

It is tempting to correlate the three pre-Wisconsin glaciations in the La Sal Mountains and Glacier National Park with the Nebraskan, Kansan, and Illinoian stages of the Mississippi Valley. Some support for this reasoning is found in the following similarities in the two regions.

(1) Only three pre-Wisconsin stages have been found in the Mississippi Valley region. Only three stages have been found in the Rocky Mountains.

(2) Buried weathered zones on each of the three tills or associated deposits in the Mississippi Valley region and in the Rocky Mountains are alike, particularly in their range of color and degree of clay accumulation, and are much more deeply decomposed than those on younger deposits.

(3) In the Rocky Mountains, the two older tills are either superposed or rest on two relatively closely spaced erosion surfaces. In a few places, they are overlapped by the youngest till. In the Mississippi Valley, three pre-Wisconsin tills are superposed disconformably in many places.

(4) A similar interglacial erosional history appears to characterize both the Rocky Mountains and the major drainage west of the Mississippi River.

(a) In the Rocky Mountains, erosion of canyon proportions both preceded and followed the third pre-Wisconsin stage.

(b) In the northwestern part of the Mississippi Valley, extensive Yarmouth dissection is believed to have preceded diversion of the Missouri River by Illinoian ice, subsequent to which, in Sangamon and later time, the river channel became entrenched (Warren, 1949; 1952; Crandell, 1953; Flint, 1955; Simpson, H. E., in preparation).

(c) In the central western part of the Mississippi Valley, along the drainage of the South Platte River in Colorado, the lowest of three pre-Wisconsin pediments lies below the more closely spaced upper two and is deeply entrenched by valleys in which are a complex of Wisconsin and younger terrace deposits (Bryan and Ray, 1940). In eastern Colorado and western Nebraska, deposits of Nebraskan and Kansan age lie high on the valley sides above terrace deposits of Illinoian age, below which, along the valley floors, are Wisconsin and Recent deposits. To the east, in central and eastern Nebraska, Illinoian deposits lie in relatively minor disconformity on late Kansan or Yarmouth deposits (Grand Island formation) in valleys that are trenched into older Kansan deposits (Red Cloud formation) on the uplands (Condra, Reed, and Gordon, 1950; Schultz, Lueninghoener, and Frankforter, 1951). The writer believes that this late Kansan dissection in central and eastern Nebraska is atypical of the Great Plains. It has been explained as the result of blocking and reopening of drainage along the margin of the continental Kansan ice sheet, together with the erosional effects of local meltwaters during its recession (Condra, Reed, and Gordon, 1950, p. 20). A similar

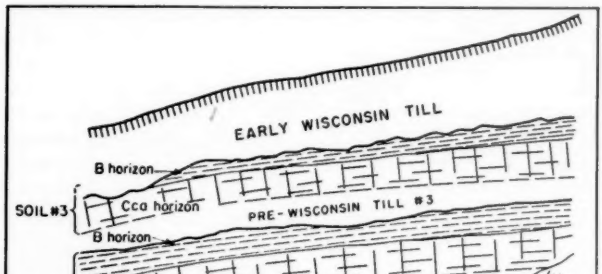

FIGURE 3.—DIAGRAMMATIC SECTION OF THREE SUPERPOSED PRE-WISCONSIN TILLS ON THE WEST SLOPE OF ST. MARY RIDGE

history is postulated for certain streams in central and eastern Kansas (Lohman, 1948; Frye and Leonard, 1952).

(d) Along the drainage of the Arkansas and Cimarron Rivers in southeastern Colorado and southwestern Kansas, superposed Kansan and Nebraskan deposits on the uplands are cut by valleys containing remnants of late Pleistocene high-terrace gravels (Latta, 1944; McLaughlin, 1946; 1949; 1954) of probable Illinoian age (W. D. Taylor, personal communication). These deposits are cut by valleys containing extensive Wisconsin and Recent alluvium. To the east, post-Kansan dissection decreases, with the result that Illinoian and Sangaman deposits—Kingsdown formation of Hibbard (1949; 1955) or Crete member of Sanborn formation of Frye and Leonard (1952)—lie either in shallow gullies that dissect Kansan deposits (Latta, 1950; Hibbard, 1955) or, more commonly, overlap those deposits.

Except for conditions along the Platte River in Nebraska that can be accounted for locally, the similarity in pre-Wisconsin erosional history of these areas suggests correlation of the three glacial stages in the mountains with those of the continental ice sheets. The erosion was probably the product of epeirogenic uplift that was greatest in the mountains and decreased eastward to negligible proportions in central Kansas and eastern Nebraska.

Opposing the suggested correlation are the following arguments:

(1) On the basis of available evidence, more than three pre-Wisconsin glaciations may be present in both the La Sal Mountains and Glacier National Park.

(2) Because the buried weathered zones on the tills are alike, it is impossible to correlate one till with another on the basis of soil-profile characteristics either within the Rocky Mountains or between the Rocky Mountains and the Mississippi Valley.

(3) Details of the erosional history of the two regions are different. For example, in Iowa, though erosion followed each pre-Wisconsin interval of soil development, greater erosion is believed to have taken place in late Aftonian time than in Yarmouth or Sangamon time (Kay and Apfel, 1929; Kay and Graham, 1943). In central and eastern Nebraska and northern Kansas, the major interval of erosion took place in late Kansan time during recession of Kansan ice and before Sangamon soil development (Condra, Reed, and Gordon, 1950; Schultz, Reed, and Lugn, 1951; Schultz, Lueninghoener, and Frankforter, 1951).

(4) Interglacial deposits, whose character, contained fauna, and relation to deposits of glacial stages provide many of the most critical criteria for correlation within the Mississippi Valley region, are lacking in the Rocky Mountains.

(5) No diagnostic fauna has been found in any deposits that can be stratigraphically correlated with the sections in which three glaciations are demonstrated.

(6) Numerical sequence is not positive evidence for correlation.

In conclusion, at least three pre-Wisconsin glaciations are evident in two areas in the Rocky Mountain region. Correlation between the two areas or with the standard section of the Mississippi Valley is plausible but not proved.

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Short Notes

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GRAVITY ANOMALIES AND STRUCTURE OF THE WEST INDIES

PART III

BY G. LYNN SHURBET AND J. LAMAR WORZEL

In Part I (Ewing and Worzel, 1954) and Part II (Worzel and Ewing, 1954) of this series, a north-south gravity profile through San Juan, Puerto Rico, was interpreted with the aid of topographic and seismic information. It was shown that the large negative gravity anomaly associated with the Puerto Rico Trench could be explained by a downward deflection (about 7 km) (Part II, p. 195) of the crust several orders of magnitude less than the deflection generally attributed to a "tectogene." In Part I (p. 171) it was pointed out that the thickness of the crustal rocks under the trench could be either slightly thicker or thinner than normal oceanic thicknesses without contradicting the gravity data or available seismic data. More recently, a slightly different structure section than presented in Part II (p. 198) has been deduced using other density assumptions (Shurbet and Ewing, 1956, p. 532); however, nothing vital to the reasoning in Part I or Part II is changed.

During April 1955, 26 new submarine-pendulum observations were completed near Puerto Rico in USS BALAO (SS 285). Nineteen stations form two profiles across the Puerto Rico Trench, and seven stations are located near the Virgin Islands. Free-air anomalies for these stations are shown in Figure 1 with previous free-air anomalies at sea in the Puerto Rico-Virgin Islands area from sources referenced in the figure. Contours of modified Bouguer anomalies on Puerto Rico[1] and simple Bouguer anomalies on St. Thomas and St. Croix (Shurbet, Worzel, and Ewing, 1956) are shown.

[1] The contours on Puerto Rico are mainly reproduced from Shurbet and Ewing (1956, Pl. 1), but small changes and additions have been made from more recent, unpublished data.

In Figure 2 are shown the free-air anomalies (circles) observed along the line A-B (Fig. 1) and the anomaly curve (solid line drawn through the computed points) computed from the structure section in Figure 2. A method of areal integration (Hubbert, 1948, p. 218), assuming the east-west dimension to be infinite, was used to compute the points on the anomaly curve relative to standard oceanic (5 km of density 1.03, 1 km of density 2.30, 4.5 km of density 2.84 overlying the mantle of density 3.27) and continental (33 km of density 2.84 overlying the mantle) crustal sections of Worzel and Shurbet (1955, p. 91). The Hubbert method of computation yields the density boundaries directly in a form more reasonable geologically than the method of approximating the structure by rectangular blocks used in Part I and Part II. Calculations show that sections determined by the two methods are in good agreement when the vertical boundaries of the block section are smoothed into reasonable geological boundaries. Arguments for assuming densities shown in Figure 2 are given in Part I (p. 170) and by Worzel and Shurbet (1955, p. 92). To illustrate the change which could result from other density assumptions, it is noted that assumption of a density of 2.50 for the sediments would lower the *M* discontinuity under the trench by less than 1 km.

Seismic stations near the gravity profile provide good control for the computations only near the Virgin Passage (Fig. 2), and here, only the thicknesses of the sedimentary layers are known. Unfortunately, in this area as well as the rest of the southern part of the profile, gravity data are sparse or nonexistent. Therefore the structural configuration shown in

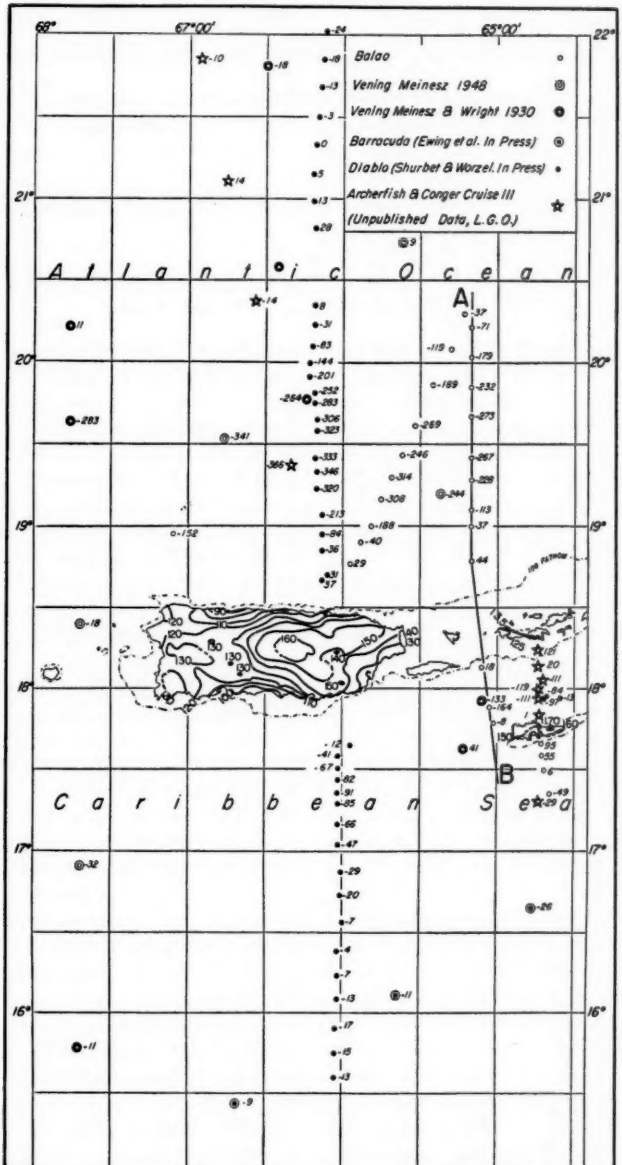

FIGURE 1.—GRAVITY ANOMALIES, PUERTO RICO-VIRGIN ISLANDS AREA

Free-air anomalies are shown at sea, and contours of modified Bouguer anomalies on Puerto Rico and simple Bouguer anomalies on St. Thomas and St. Croix are shown.

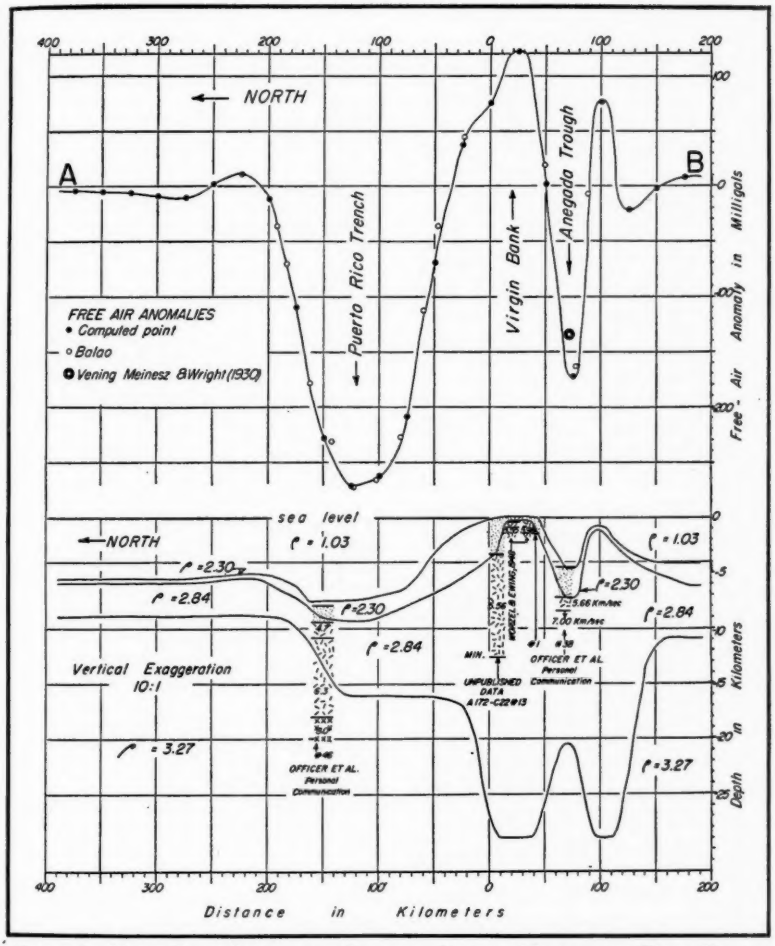

FIGURE 2.—NORTH-SOUTH GRAVITY AND STRUCTURE PROFILES THROUGH THE VIRGIN PASSAGE
Seismic sections and densities assumed for gravity calculations are indicated.

Figure 2 to the right of the zero distance point is essentially reproduced from earlier, more-detailed gravity analyses at locations to the west across Puerto Rico (Shurbet and Ewing, 1956, p. 532) and to the east across St. Thomas and St. Croix (Shurbet, Worzel, and Ewing, 1956, p. 1534). It can be said only that the gravity measurements over this area do not contradict such a configuration.

Only one seismic section is known near the gravity profile over the Puerto Rico Trench (Fig. 2; C. B. Officer *et al.*, personal communication). Here, if we accept for the water depth a mean of soundings from several sources varying up to 0.6 km (equivalent to about 30 mgal in gravity) and the sedimentary thickness determined by Officer *et al.*, the depth to the *M* discontinuity necessary to satisfy the observed gravity anomalies is about 3 km less than the depth determined seismically, a discrepancy about twice the seismic observational errors and equivalent to about 30 mgal in gravity.

This discrepancy is not serious, considering that the water depth is not known with sufficient accuracy to permit more nearly perfect agreement between the seismic and gravity results.

Thus the conclusion of small downward deflection of the crustal rocks in the Puerto Rico Trench presented in Part II (p. 195) is confirmed in Figure 2, although the section is not well controlled seismically. Some thickening of the crustal rock section under the trench from normal oceanic thicknesses is shown by the seismic data (C. B. Officer *et al.*, personal communication) and confirmed by gravity calculations.

Valuable assistance in arranging and conducting the cruise to make these measurements was given by the Commander Submarine Force, Atlantic Fleet, and the officers and crew members of USS BALAO. It was largely through the efforts of LCDR. W. J. Flannery, Commanding Officer, USS BALAO, that sufficient time was available to make these measurements. K. Gilchrest assisted the first author in the observing program, B. Skinner assisted in the computations, and A. Trefzer made the drawings.

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FROSTING OF QUARTZ GRAINS BY CARBONATE REPLACEMENT

BY THEODORE R. WALKER

Frosting on quartz grains is commonly attributed to abrasion during aeolian transportation. The effectiveness of sand blasting in producing frosted surfaces cannot be questioned, but not all frosting originates in this way. A review of the literature dealing with surface textures of quartz grains reveals several other explanations of frosting, including differential solution of grain surfaces by percolating ground waters, incipient quartz overgrowths, and pressure solution along contacts between adjacent grains. Still another cause of frosting, and one that has received little mention in the literature, is that caused by carbonate replacement of quartz along grain boundaries.

The writer has studied thin sections and insoluble residues of several quartzose limestones and dolomites and finds that calcite and dolomite replacement of quartz is common and that microetching on many peripherally replaced quartz grains produces a surface texture that appears frosted (Pl. 1, figs. 1–4). Frosting of this type is easily mistaken under the binocular microscope for that resulting from aeolian abrasion, and failure to interpret its origin correctly may lead to erroneous conclusions concerning genesis of these carbonate sediments.

Misinterpretation of replacement frosting is most likely to result if the peripherally replaced grains are well rounded, because such grains have what might be considered the ideal shape of wind-transported sand grains. Partially replaced grains in most samples observed, however, are not well rounded because differential penetration of quartz by the replacing carbonates tends to increase grain angularity. The origin of the grain frosting on angular grains might be puzzling to the observer, but it probably would not be attributed to aeolian origin. The writer has seen many examples, on the other hand, in which replacement apparently had only slightly modified the shape of grains that were well rounded when deposited, and these rounded and frosted grains were readily mistaken for aeolian material (Pl. 1, figs. 2, 4).

Frosting observed on grains in most samples is a result of simple marginal replacement of quartz. The shape of the relic grains therefore depends in part on the shape of the original detrital grains and in part on the degree to which carbonates differentially penetrate the grains. Well-rounded and frosted quartz grains of more complex origin, however, have been observed in samples of oölite from the Ordovician Oneota formation near Madison, Wisconsin. The nuclei of the oölites in these samples are well-rounded quartz grains, many of which have secondary quartz overgrowths. The silica of the overgrowths in these samples is more susceptible to replacement than is the silica of the original quartz grains, and replacement tends to halt at the grain boundaries. Selective replacement has removed the overgrowth quartz on most grains and has produced frosted relics which have the well-rounded shapes of the original grains. These grains resemble frosted grains of aeolian origin (Pl. 1, figs. 3, 4). Steps in the replacement that has produced these unusual grains are shown in Figure 5 of Plate 1.

When considering the mechanism of quartz replacement by carbonates, two questions immediately come to mind. To what extent can replacement destroy detrital quartz grains in carbonate sediments, and what happens to the silica that is replaced?

In extreme examples quartz grains in carbonate sediments locally might be replaced completely and might result in a carbonate rock free of detrital quartz. Evidence of such replacement may be preserved as relic outlines of original grain boundaries, or it may be destroyed during the process of replacement or by later recrystallization. Complete replacement of detrital quartz is probably more significant than has been suspected.

The problem of what happens to the replaced silica is important because it concerns the problem of the origin of secondary silica in

some sediments. An appreciable amount of silica is released by even partial replacement of quartz, and in those rocks in which replacement has been extensive it furnishes a volumetrically important source of material for secondary silification in associated sediments. After replacement, the silica may not travel far before it encounters an environment favorable for its precipitation. Such an environment might be a change in pH of the migrating solutions which would result in silica deposition by direct precipitation or by replacement (Correns, 1950, p. 53; Newell *et al.*, 1953, p. 162–166). The writer believes carbonate replacement of quartz (and also feldspar) as a local source of silica for secondary silicification has not received the consideration it deserves. In the forthcoming paper the writer will discuss this subject in detail.

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Plate 1.—CARBONATE REPLACEMENT OF QUARTZ

Figure 1.—Dolomite replacement of quartz-grain boundaries. Original well-rounded shape of quarts has been only slightly modified by replacement. Williams Canyon dolomite (Devonian?) near Manitou Springs, Colorado. Plain light, 50×.

Figure 2.—Insoluble residue of sample shown in Figure 1 of this plate. Frosting on grains is due to marginal replacement by dolomite. Sample has been sieved for clarity of illustration; fraction shown is medium sand grade. 15×.

Figure 3.—Selective replacement of overgrowth quartz by dolomite. Dark rings are oölite shells; dark spots are tangential sections. The oölite nuclei were originally surrounded by secondary quartz overgrowths that have been replaced by dolomite (gray mosaic within dark rings). The sequence of this replacement is shown in Figure 5 of this plate. Oneota dolomite (Ordovician) near Madison, Wisconsin. Plain light, 50×.

Figure 4.—Insoluble residue of sample shown in Figure 3 of this plate. Frosting on grains is due to marginal replacement by dolomite. Well-rounded shapes reflect rounding of quartz nuclei in Figure 3. A grain showing a remnant of unreplaced quartz overgrowth is near the center of the photo. Sample has been sieved; fraction shown is medium sand grade. 15×.

Figure 5.—Sequence of selective replacement of secondary quartz overgrowths. Roundness of overgrowths probably is due to abrasion, and indicates multiple cycle sand grains. Oneota dolomite (Ordovician) near Madison, Wisconsin. Plain light, 50×. A. Initial replacement of quartz overgrowths by dolomite. Note that replacement tends to halt at boundary of nucleus. B. Intermediate stage of overgrowth replacement. Dolomite has penetrated the nucleus at lower left but, in most places, has halted at the nucleus boundary. Adjacent clear areas are secondary silica that predates the dolomite. C. Advanced stage of overgrowth replacement. The overgrowth is almost completely replaced by dolomite. Unreplaced remnants remain along the left and upper right sides of the nucleus. Slight penetration of the nucleus is visible at lower left. Further replacement would remove all traces of the secondary overgrowth, and differential penetration of the nucleus would continue.

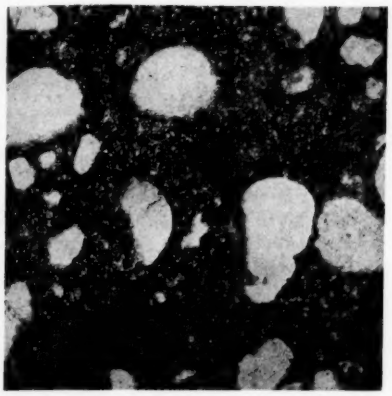

FIGURE 1

FIGURE 2

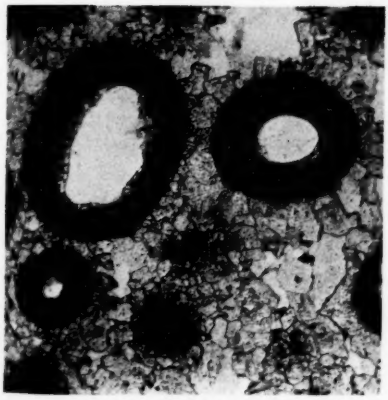

FIGURE 3

FIGURE 4

FIGURE 5A

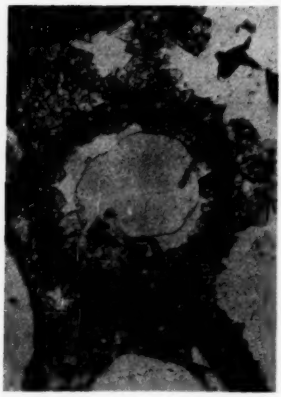

FIGURE 5B

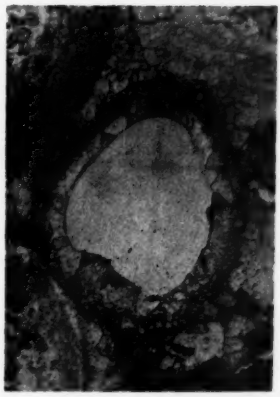

FIGURE 5C

CARBONATE REPLACEMENT OF QUARTZ

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FAULTS AND MONOCLINES ON KILAUEA VOLCANO, HAWAII

BY GORDON A. MACDONALD

High-angle normal faults are numerous on Kilauea—around the caldera, bounding grabens along the rift zones, and in a zone parallel to the south coast of the island along which the central part of the volcano has been elevated in relation to the portion farther south. In the Hilina fault system, near the south coast (Stearns and Macdonald, 1946, p. 40, 129–130; Wentworth and Macdonald, 1953, p. 21), the scarps are heavily mantled by lava flows, and the nature of the displacement rarely can be determined. Around the caldera and along the rift-zone grabens, the fault scarps commonly pass into monoclines. In the caldera area a mantle of pyroclastic debris several feet thick covers the lavas, but at many places cracks and gullies reveal the monoclinal flexure in the lava flows beneath the pyroclastic rocks. The best-exposed scarps lie principally along the southern edge of the graben along the southwest rift zone of the volcano (Stearns and Macdonald, 1946, p. 129) and in the zone of faults that diverges eastward from the southwest rift zone and passes 2–3.5 miles south of the caldera. Excellent examples of all the features described can be found in the latter fault zone within 3 miles on each side of the Hilina Pali road. (For maps showing the location of the road and the fault zone *see* Stearns and Macdonald, 1946, Pl. 1; Finch and Macdonald, 1953, Pl. 2.)

In a general way the scarps are curvilinear in plan, but in detail they are zigzag, with sharp salient and re-entrant angles.

The monoclines range up to about 50 feet in height, although most are less than 30 feet. The slope of the risers of the monoclines generally is between 30° and 50° but locally is less than 10° and greater than 75°.

Surficial flow patterns on the lavas indicate that the lava flows forming the surface of the monoclines moved in a direction diagonally across, or even directly opposite, the present slope of the scarps, and vents from which the flows issued are on the downthrown side of the scarps. Thus the monoclines cannot be the result of mantling of fault scarps by flows. Unquestionably, they were developed in the thin-bedded lavas after consolidation of the flows. It may be a little surprising that such brittle rocks as basaltic lava flows can undergo flexuring at the surface, but the bending is in places smooth and regular. It appears to be accomplished largely by slight opening and closing of the numerous nearly vertical joints in the lavas, together with some slippage between beds. The slippage is facilitated by the loose clinkery tops and bottoms of aa lava flows (Wentworth and Macdonald, 1953, p. 57–61).

Unbroken monoclines (Fig. 1, *A*) have been found at only three localities. The commonest imperfection in the flexure is a prominent crack following the zone of tension along the upper (anticlinal) bend (Fig. 1, *B*). The open wedge-shaped crack locally may be as much as 5 feet across and 10 feet deep.

Theoretically, a zone of surface compression should lie along the lower (synclinal) bend. On the monoclines of Kilauea this zone of compression commonly is marked by a crushed zone and low ridge along the base of the scarp (Fig. 1, *C*). Locally the ridge consists of a sharp anticlinal buckle, essentially unbroken except for an open crack along its crest, but more commonly the ridge is thoroughly brecciated. The breccia ridge appears to have been squeezed up out of the crushed zone. The ridges range in height up to 2 or 3 feet. Generally they lie directly along the base of the scarp, but in some places they diverge from it and traverse the lower slope of the scarp itself, or rarely the nearly flat lava beds immediately adjacent to the scarp.

The breccia consists principally of angular fragments of the adjacent flow rock, from 1–2 inches to a foot in diameter. Fragments larger than a foot across are uncommon, and surprisingly little fine matrix is present. However, the fine material probably has been partly, or even largely, removed by rain wash.

The breccia ridges resemble some pressure ridges in lava flows (Wentworth and Mac-

donald, 1953, p. 49) and indeed were formed in a somewhat similar manner; but their peculiar spatial relationships to the scarps, plus the absence of pressure ridges on adjacent parts of the flow surface, indicate that they are not ordinary pressure ridges. The ridge of crush breccia along the synclinal bend of monoclines does not appear to have been recorded elsewhere, probably because the broken material could be rapidly removed by erosion. Its preservation on Kilauea volcano is the result partly of the recency of the deformation, and partly of the unusually high permeability of the surface rocks which results in very slow erosion either by sheet runoff or by streams in early stages.

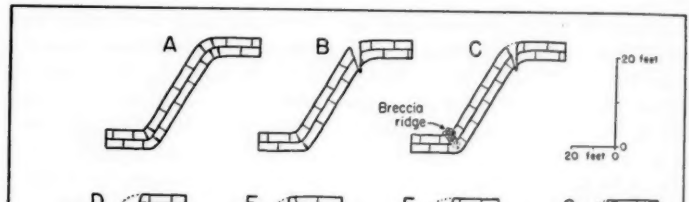

FIGURE 1.—DIAGRAMS ILLUSTRATING VARIATIONS IN MONOCLINES AND FAULTING AT KILAUEA VOLCANO. The vertical scale is exaggerated.

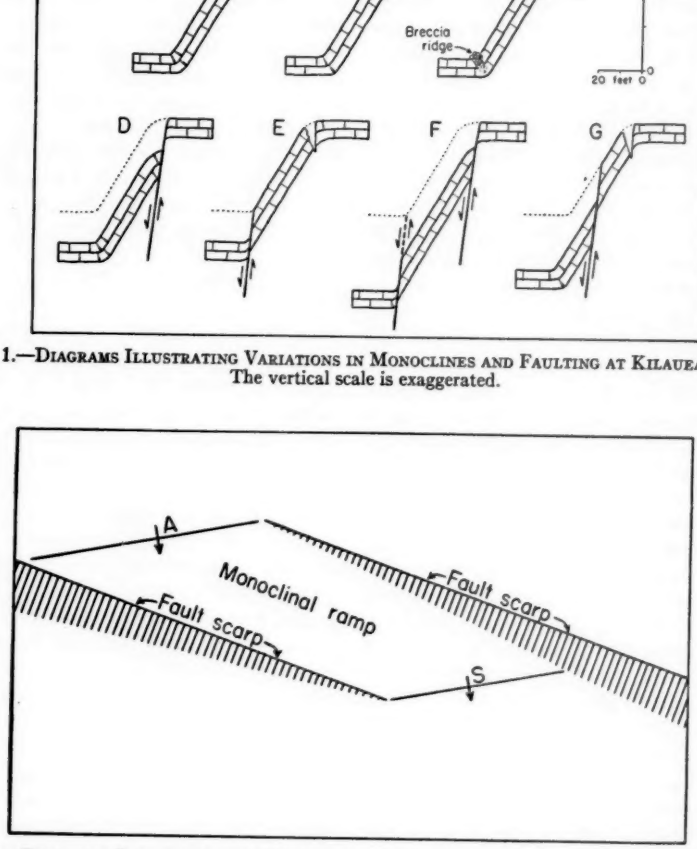

FIGURE 2.—DIAGRAMS ILLUSTRATING A MONOCLINAL RAMP BETWEEN TWO ECHELON PORTIONS OF A SPLINTERED FAULT SCARP.

Commonly the monoclines are broken by faults. The most usual position for the fault scarp is at the top of the monocline (Fig. 1, *D*), probably because the anticlinal portion of the flexure is weakened by the tendency to slit open. At some places the upper edge of the riser of the monocline rests tightly against the fault scarp (Fig. 1, *D*), but at many places

there is an open wedge-shaped crack like that formed at the anticlinal bend of unfaulted monoclines. Almost as commonly the fault scarp is at the base of the monocline (Fig. 1, *E*), and at other places it may lie at any position on the slope of the monocline (Fig. 1, *G*). Rarely, faults occur at both the top and bottom of the monocline (Fig. 1, *F*).

At several localities, after following the upper edge of the monocline for an appreciable distance the fault diverges from the anticlinal bend and extends diagonally across the monocline, then continues along the synclinal bend. At any point along the diagonal portion of the course of the fault, a vertical cross section would appear somewhat like that of Figure 1, *G*. At some localities the fault trace merely diverges from the top or base of the flexure, leaving the anticlinal or synclinal bend unbroken. More commonly, however, its course along each bend is continued by a minor branch that gradually dies out along the bend.

Splintered fault scarps, resembling those described by Davis (1903, p. 30) and Cotton (1919), are fairly common along the southern edge of the graben of the southwest rift zone and its eastward extension. They consist of two fault scarps or combinations of faults and monoclines en echelon, the throw on one gradually decreasing as that on the other increases. Between the parallel portions of the two scarps lies a gently sloping monoclinal ramp (Fig. 2). At several places such ramps have been utilized for the routes of roads and trails crossing the scarps. In Figure 2 the symbols used to indicate the anticlinal (A) and synclinal (S) bends at the top and bottom of the monocline are those suggested by Kelley (1955, p. 792).

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